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SOIL HYDROLOGICAL PROPERTIES AFFECTING DRAINAGE IN THE FALHER REGION

by

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A THESIS

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The undersigned certify that they have read, and recommend to the Faculty of Graduate Studies and Research, for acceptance, a thesis entitled...Soil Hydrological Properties.. Affecting Drainage in the Falher Region ....

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submitted by.....Austin Charles J. Sichinga.................. in partial fulfilment of the requirements for the degree of Master of..Science..................

ABSTRACT

The Peace River region of Alberta is characterized by short frost-free growing seasons. High soil moisture contents result in trafficability problems which may delay seeding and harvesting and, because of the short crop seasons, cause crop losses.

The drainage problem in the Peace River region has been attributed to a combination of the large volumes of water from snowmelt during seeding, high precipitation while evapotranspiration is low during harvest and soil hydrologic properties. In this study soil physical properties were quantified. Water entry into, and its flow through, the soil and soil moisture status during the 1981 seasone, were investigated at three sites.

The analysis of soil physical properties showed that clay contents in these soils are high, above 50% and sand contents low, below 10%, with clay contents highest in the B horizon. Bulk densities were also found to be highest in the B horizon. Saturated hydraulic conductivities of these soils were found to be very low in the Ap horizon and even lower in the B horizon. Flow of water through these soils was found to be largely controlled by the B horizon. A comparison of the three sites showed site 3 to have a much less pronounced difference in physical properties between the Ap and the B horizon.

The initial rate of water entry into the soil was found to be high when the soils were dry and cracked. The rate of water entry into the soil was found to quickly decrease when water penetration

reaches the B horizon, which acts as a barrier to water percolation. Further addition of water results in increased moisture content in the Ap horizon and subsequent surface ponding.

Moisture changes during the season were found to be mainly restricted to the top 50 cm of the soil. Changes in soil moisture contents were found to be influenced by a number of factors including evapotranspiration, precipitation and the kinds of crops grown.

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INTRODUCTION

LOCATION

The Peace River District is located approximately 400 km northwest of the city of Edmonton (Figure 1). The area covers parts of the provinces of Alberta and British Columbia. The region is bounded by the Rocky Mountains to the west and the south and to the east by the highlands of the lesser Slave Lake. Elliot (1974) reports that of the over 16 million hectares in the district, 2 million of the 8 million potentially arable hectares are cultivated.

CROPS GROWN

Records from the Alberta Agriculture Statistics Branch show that in the 1979 cropping season, cereals accounted for 55% of the total land under cultivation, forages accounted for 13.6% and other crops accounted for 0.7%. Summerfallow accounted for 29.4 of the land while 12.7% was classified as new breaking land (Table 1).

CLIMATE

The geographic position of the Peace River region makes it a unique agricultural region. The limits of agricultural production have been extended, so that this area constitutes the northernmost agricultural frontier in Canada (Hoyt et al., 1974). A combination of factors, mountains to the west, which cut off cool air masses and extra long summer days, a consequence of latitude, result in unusually warm summers which make possible

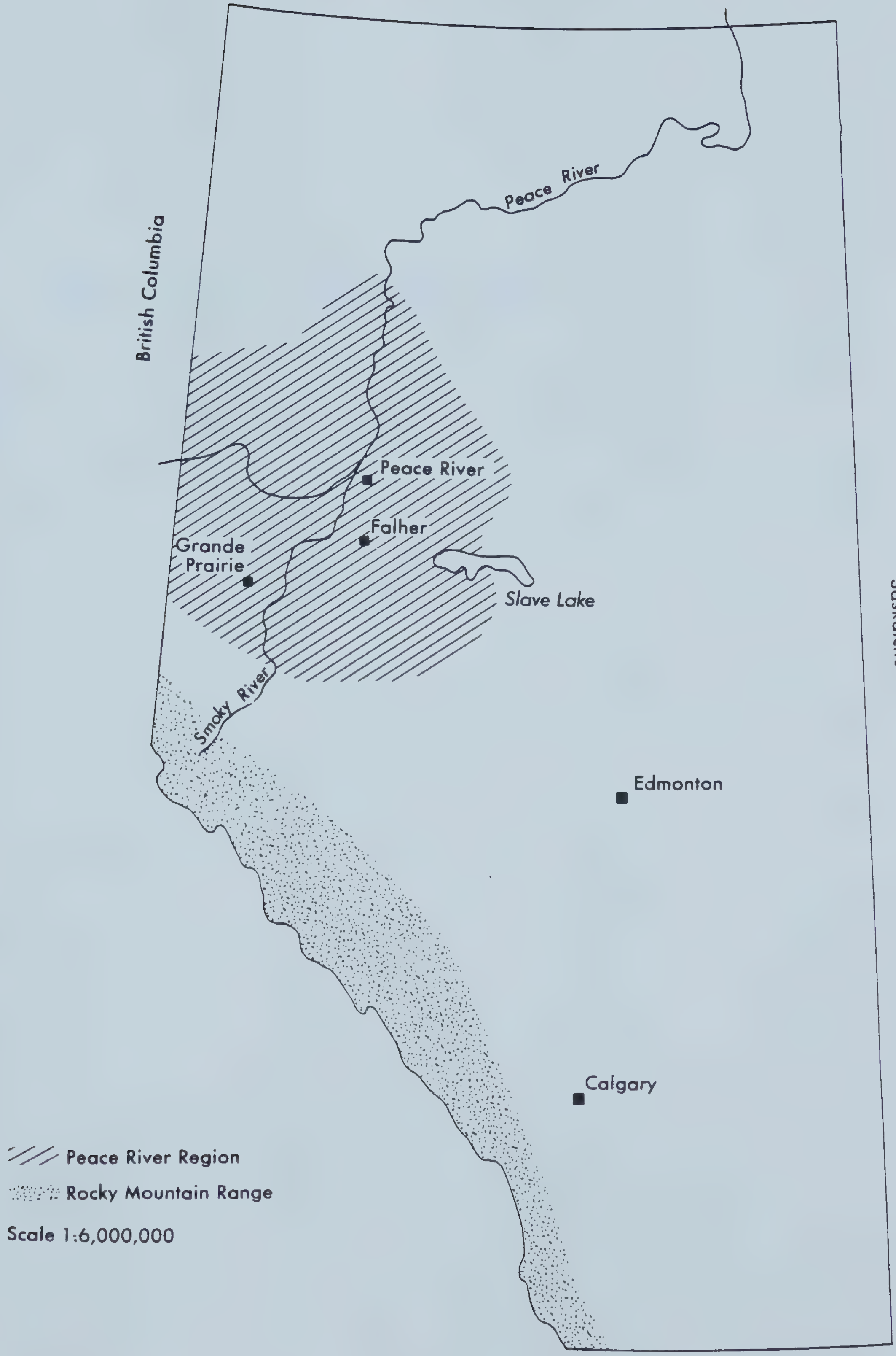

Figure 1 Location of the Peace River Region

TABLE I

CROP AREAS FOR THE PEACE RIVER REGION 1979
(after Faris et al., 1981)

Crop	Hectares in 000's	% of Total Cultivated Land
Rape Seed	477	26.5
Barley	307	17.0
Wheat	142	7.9
Oats	54	3.0
Flax	8	0.4
Rye	4	0.2
Cereals (Total)	992	55.0
Forages	245	13.6
Other Crops	12	0.7
Summerfallow	531	29.4
New Breaking	23	1.3
Total Cultivated	1803	100

the extension of the agricultural frontier (Richards, 1968).

The Peace district is characterized by short summer seasons and relatively short frost-free growing seasons. The use of early maturing crop varieties and strict adherence to recommended early seeding dates is thus essential for successful crop production in the region. Factors which interfere with or delay field operations are therefore of major concern to the farming community in this region.

It is with respect to the foregoing discussion that the significance of the drainage problem should be viewed in the Peace district. Snowmelt in spring and high rainfall during the harvest period coupled with low evapotranspiration rates during these periods may result in surface ponded water and excessively muddy fields, delaying seeding and harvesting in spring and late summer respectively.

The drainage problem is not unique to the Peace River region. The problem is encountered in flat lands whenever precipitation rates are higher than rates of water movement through the soil. However, even though the drainage problem occurs elsewhere, the causes and hence solutions to it are varied and will depend on the unique characteristics of the climate, soils and crops for each specific region.

Therefore, even though studies and management recommendations have been made on the subject for areas such as southern Alberta, these studies and recommendations may not necessarily be relevant to the Peace country. A thorough investigation of the climate and physical characteristics of the soils of the Peace River district is, thus, a necessary first step towards finding and recommending solutions to the

drainage problem.

RESEARCH OBJECTIVES

The objective of this research, therefore, was to identify, through an indepth study, soil physical characteristics which may contribute to drainage problems in the Falher area of the Peace River district.

To realise the objective, specific investigations undertaken may be summarised as follows:

(i) precipitation and evapotranspiration comparisons,

(ii) water infiltration into and redistribution within the soil profile,

(iii) soil horizonation and differences in bulk densities and the effect of these on water movement through the profile, and

(iv) soil water holding capacities and profile moisture contents during the cropping season.

LITERATURE REVIEW

INTRODUCTION

The word drainage is often used to refer to the physical network of streams and surface waterways in a given area or to the water carried in these streams (Luthin, 1957). In this study, however, the word drainage is used to mean the removal of excess water from wet agricultural land (Schwab *et al.*, 1966).

Soil water is considered excessive if it is more than that required for optimum crop growth or if it does not allow for the efficient use of agricultural machinery without damage to the machinery or to soil structure. Excessive soil moisture conditions arise when a high percentage of soil voids are water filled, leaving only a few voids air filled. Artificial drainage becomes necessary whenever soil moisture is excessive and when water is ponded on the soil surface because not enough water is removed from the soil naturally (Hillel, 1980b).

Generally drainage problems arise when water addition to the soil exceeds water loss from the same, such that an optimum soil-water soil-air balance is not maintained. Drainage problems, therefore, arise when precipitation, surface and subsurface water flow into a given field exceed evapotranspiration, deep percolation and surface runoff of water from the same field i.e. water balance between the inflow, I, and the outflow, O, is not maintained and storage, S, is increasing. This relationship can be described by the following equation:

$$I - O = \frac{dS}{dt} \quad \text{where } t \text{ is time.}$$

Drainage problems can be attributed to a number of environmental and/or soil profile conditions. These conditions can be summarized as follows:

(i) Low rates of evapotranspiration which reduce total water loss from the soil.

(ii) Low rates of infiltration and/or high intensity precipitation, such that water enters the soil at a rate lower than that at which it is being added to the soil surface, resulting in ponded water.

(iii) Low rates of pecolation. Water flows downward through the soil at such a slow rate that excess moisture accumulates.

(iv) The presence of a high water table which increases soil moisture content of the near surface horizons through capillary rise.

(v) Low position of the land in the landscape of the area which may result in the field acting as a seepage zone as well as a surface drainage trough. The low topographic position may also reduce chances of natural drainage outlets for ponded surface runoff water.

Adequate drainage provides the maintenance of optimum soil moisture conditions, maintaining a balance between incoming water, basically infiltration, and outgoing water, including moisture redistribution and percolation. The maintenance of the water balance is predominantly

governed by water movement into and through the soil.

FACTORS THAT CAUSE AND GOVERN WATER MOVEMENT

Water Potential. Water, like all matter on earth, possesses energy (Hillel, 1980a).

Water possesses energy in two forms, as kinetic and as potential energy. The kinetic energy of water is that energy which the water possesses owing to its movement. This energy can be represented by the formula:

$\frac{V^2}{2g}$ where V is velocity and g is acceleration due to gravity.

The potential energy of water is that energy which is due to either pressure differences or elevation differences (Hillel, 1980a;Hansen et al., 1980). The combined kinetic-potential energy, H, may be represented by Bernoulli's equation which gives energy per unit mass as:

$$H = \frac{V^2}{2g} + \frac{P}{W} + y$$

where P is pressure per unit area, W is weight of water per unit volume, and y is elevation above some chosen datum.

In the flow of water through soils, since the velocity of flow is usually very low, the kinetic energy $\frac{V^2}{2g}$ must be even smaller and can thus be ignored. Bernoulli's equation can thus be simplified to:

$$h = \frac{P}{W} + y$$

Where h is the hydraulic head, also called the Piezometric head.

At the surface of water which is open to the atmosphere, the hydrostatic pressure is zero. Therefore water in a saturated profile is at hydrostatic pressure greater than atmospheric and is said to possess a positive pressure potential. Water in an unsaturated profile is at a hydrostatic pressure less than atmospheric and is said to possess a negative pressure (also called tension or suction).

The hydrostatic pressure, P, of water is related to the hydraulic head, h, by the equation:

$h = \frac{P}{\delta g}$ where δ is water density and g is acceleration due to gravity.

In the soil, water movement is in response to differences in this hydraulic head, h. The rate at which water moves is dependent on the hydraulic gradient and the hydraulic conductivity.

Darcy's Law. An observation first made and reported by the French engineer Henri Darcy in 1856 showed that the discharge rate, Q, equalled the volume, V, flowing through a column of length L, per unit time, t. It was directly proportional to the cross sectional area, A, and to the hydraulic head drop ΔH, and was inversely proportional to the length over which flow occurred.

The relationship between the hydraulic gradient, $\frac{\Delta H}{L}$, the hydraulic conductivity, k, and the rate of discharge is best described by Darcy's Law given by:

$Q = \frac{V}{t} = \alpha \; A \; \frac{\Delta H}{L}$ where α is the proportionality factor.

The specific discharge rate $(\frac{Q}{A})$, called flux density or flux, q, can be shown to be proportional to the hydraulic gradient:

$$q = \frac{Q}{A} = \frac{V}{At} = \alpha \ \frac{\Delta H}{L}$$

The proportionality factor, α, is given the symbol k and is the hydraulic conductivity.

Thus $q = k \ \frac{\Delta H}{L}$

This law states that flow of liquid through a porous media is in the direction of and at a rate proportional to the driving force, ΔH, acting on the liquid and proportional to be hydraulic conductivity.

Hydraulic Gradient. The hydraulic gradient is difference in hydraulic head between two points divided by the distance between these points. It is the head drop per unit distance in the direction of flow and is the driving force for water movement (Hillel, 1980a). Water will not move unless a hydraulic gradient exists. It will continue moving until the hydraulic gradient drops to zero, when hydraulic head equilibrium has been reached.

Hydraulic Conductivity. The time it takes for hydraulic head equilibrium between two points to be reached is determined in part by the hydraulic conductivity of the soil.

The hydraulic conductivity is a ratio of the flux or rate of discharge to the hydraulic gradient. Hydraulic conductivity is influenced by a number of factors which include soil and fluid factors. The soil factors include texture, structure, density, total porosity, pore sizes and pore geometry. The fluid attributes that influence hydraulic conductivity are viscosity and density.

O'Neal (1949) found that fine-textured, platy and prismatic structured

soils have low hydraulic conductivities. He also found that it is not only the structure, but also the relationship between the length of vertical and horizontal axes of the structural aggregates that are influential in hydraulic conductivity. The overlap of these aggregates is also important, the hydraulic conductivity being higher in the direction of the overlap.

Texture also affects hydraulic conductivity. High clay and silt percentages reduce hydraulic conductivities as do high densities (Mason *et al.*, 1957; O'Neal, 1949). Saturated hydraulic conductivities are likely to be higher in soils that are cracked and where macropores make up the greater percentage of total porosity.

If the hydraulic conductivity of a soil varies from point to point, the soil is said to be nonhomogeneous. The soil is said to be anisotropic if the hydraulic conductivity varies with direction at a point. Hydraulic conductivity also varies with soil water content, being higher at higher moisture contents, and is also affected by hysteresis, whereby the soil wetness, at a given suction, depends on whether the process leading up to it was absorption or desorption.

INFILTRATION

Introduction. Infiltration rate is defined as the time rate at which water enters the soil (Schwab *et al.*, 1966). Infiltration rate is also called intake rate where infiltration occurs under a specific soil surface configuration (Israelsen and Hansen, 1962). Although infiltration has sometimes been defined to mean not only water entry but also the associated downward flow of this water through the profile,

infiltration in this study is considered as merely the entry of water into the soil; the downward flow of this water is considered redistribution or percolation.

Infiltration capacity is the maximum infiltration rate for given soil conditions. Horton, who in 1933 introduced this concept, found that infiltration capacity is highest when water is first added to the soil and that it drops with time until it reaches a constant value (Arend and Horton, 1942). This final constant infiltration rate has been called basic intake rate or steady state infiltrability (Hillel, 1980b).

Accumulated intake is the total amount of water absorbed by the soil for a given period of time since the onset of infiltration.

Factors that Influence Infiltration Capacity. Initial infiltration capacity of the soil is dependent on a number of soil properties. These properties include initial soil moisture content, the hydraulic conductivity of the soil, the surface conditions of the soil and the presence within the profile of less permeable flow impeding layers.

Initial soil moisture affects soil infiltrability through its influence on soil hydraulic gradient. An increase in initial soil water content is accompanied by a reduction in the hydraulic gradient. Low initial moisture contents will thus give relatively higher initial infiltration capacities.

The hydraulic conductivities of the surface and the subsurface horizons determine the maximum rate of water flow through the profile. Soil properties, that influence hydraulic conductivity, therefore, are

important in influencing infiltration. Soil porosity, pore geometry, structure and texture are important as are soil and water temperatures. Low saturated hydraulic conductivities indicate low potential infiltration rates.

Soil surface conditions also influence infiltrability. Well granulated, vegetated and highly porous soil surfaces generally have higher infiltrabilities. Disturbed, bare and compact soils generally have lower infiltrabilities (Duley, 1939). Soil surface crusts can act as hydraulic barriers and impede infiltration, therefore, soils with unstable structures which tend to form crusts have lower initial and final infiltrabilities (Diebold, 1954; Hillel, 1980b).

Layering within the soil profile also affects infiltration. Whenever layers of low hydraulic conductivity are close to the soil surface, they may affect infiltrability soon after its onset. If these layers occur deeper in the soil profile, they may only be significant in influencing the final infiltrability.

Factors Responsible for the Drop in Infiltrability. The characteristic drop in infiltrability from an initial high to a constant low final infiltrability has been shown by a number of investigators (Horton, 1933; Arend and Horton, 1942; Bodman and Colman, 1943). This drop has been attributed to a number of processes that occur during infiltration. These processes can be summarized as follows:

(i) A drop in the hydraulic gradient due to the decrease in the soil suction which occurs as more and more water enters the soil and as the wetting front moves deeper into the soil profile (Bodman and Colman, 1943).

(ii) The gradual compaction and crusting of the soil surface under the impact of rain drops. The resulting crust has a lower permeability and acts as a barrier to infiltration (Duley and Kelly , 1939; Horton, 1941; Moore, 1980).

(iii) Soil slaking, detachment, migration and inwashing of fine soil particles as a result of rain drop impact. The inwashing of the fine soil particles leads to blocking of water conducting pores and hence lower infiltrability (Horton, 1941; Moore, 1980; Hillel, 1980b).

(iv) The expansion of 2:2 and 2:1 clays such as montmorillonite and smectites in the soil which results in fewer conducting pores and thus lower soil hydraulic conductivity.

(v) The presence and compression of air trapped below the wetting front which reduces the total percentage of water conducting pores and hence lowers infiltrability (Baver, 1937).

Significance of Infiltration in Drainage Studies. Infiltration separates the portion of a given precipitation that enters the soil and that which ends as ponded water or surface runoff (Arend and Horton, 1942). In drainage studies, therefore, infiltration capacity can be used to predict whether a given intensity of precipitation will create surface ponded water and/or runoff or whether all the precipitation water will enter the soil quickly enough and not create ponded water conditions. Thus infiltration capacity is an indication of the maximum precipitation rate that will not create drainage problems.

Steady state infiltrability or final infiltration capacity may be used as an indicator of possible low redistribution and percolation rates since low steady state infiltrability is an indication of a low rate of water movement through the profile.

REDISTRIBUTION AND PERCOLATION

Introduction. Redistribution is the movement of infiltrated water within the profile. It may involve lateral and or vertical movement of water. Redistribution occurs in soils in which the hydraulic head is not uniform.

Percolation is the movement of infiltrated water through a profile, beyond the relevant depth or down to the water table. Hillel (1980b) called this process internal drainage. Percolation occurs when enough water has infiltrated into the soil that the moisture content of the relevant depth is higher than that of greater depths. The process involves the flow of water either out of the relevant depth (deep percolation) or down to the water table in instances where it is within the relevant depth.

Factors that Influence Redistribution. In redistribution, infiltrated water moves from the upper wetted horizons to the underlying unwetted horizons. Redistribution is a time dependent process that is influenced by initial soil moisture, the wetting depth of infiltration, the relative moisture contents of the unwetted layers and the water conductive properties of the soil.

In redistribution, because of the unsaturated deeper layers, the water's driving force consists of both the gravitational potential and the suction gradient of the unwetted layer. Since the suction gradient is dependent on soil moisture content, the drier the unwetted layers and the smaller the wetted depth, the greater the suction gradient and therefore the faster the redistribution process (Peck, 1970).

With the movement of water from the wetted to the drier layers in the profile, there is a decrease in both the suction gradient exerted by the dry layers as they absorb water and the hydraulic conductivity of the wetted zones as the water content of these decreases. As a result of the decrease in suction gradient and hydraulic conductivity, the flux of water drops rapidly in redistribution. The rate of advance of the wetting front will also decrease with the decrease in flux.

The redistribution process involves desorption and absorption of water, by the wetted and unwetted layers respectively and consequently is affected by hysteresis. As a result of hysteresis, the expected moisture distribution profile, in which moisture increases with depth because of gravity, is not usually observed. Hysteresis instead results in a profile in which the initially wetted, desorbing layer remains at higher moisture contents (Rubin, 1967).

In redistribution, therefore the decreasing hydraulic conductivity coupled with the decreasing hydraulic gradient and the hysteresis effect lead to higher water storage in the upper desorbing horizons.

Rubin (1967), Youngs (1960), Bresler *et al.*, (1969) and Schofield (1935) have shown that hysteretic moisture movement is slower than either the absorbing or desorbing nonhysteretic moisture movement.

Hillel (1980b) has summarized the factors that influence redistribution as those that affect hydraulic conductivity, i.e. texture, clay type and organic matter, the depth of wetting, moisture content, the presence of water flow impeding layers within the profile and evapotranspiration.

Factors that Influence Percolation. Water movement in percolation, because it occurs in near/or saturated profiles,is predominantly in response to a gravitational potential, since the suction potential is negligible.

A number of mathematical expressions have been developed to describe percolation. One presented by Hillel (1980b) assumes the profile drains uniformly and also assumes there is no flow through the soil surface. The flux q, through any plane at depth Z_b must therefore equal the rate of decrease of total water w, where $w = \theta Z_b$

$$q_b = k(\theta) = - \frac{dw}{dt} = - Z_b \frac{d\theta}{dt}$$

The downward flux increases in proportion to depth. Flux also diminishes in time as does the rate of decrease of soil wetness, in accordance with the functional decrease of hydraulic conductivity with remaining soil wetness. This equation can be used to determine hydraulic conductivity as a function of wetness where evapotranspiration is insignificant.

Initially owing to the higher saturated hydraulic conductivity, sandy soils drain much more rapidly than do the low hydraulic conductivity clays; however, in the later stages of the process, this order is reversed.

The downward flux in percolation has also been shown to be proportional to the total amount of water remaining in the profile:

$$-\frac{dw}{dt} = \lambda w$$

where $-\frac{dw}{dt}$ is a time rate decrease in profile water content w, and λ is the proportionality constant. The equation can be intergrated to give:

$$\frac{w}{w_i} = e^{-\lambda t}$$ where w_i is initial water content

This equation assumes that w tends to zero which makes it unrealistic. One must therefore allow water content to approach asymptotically some finite value of water retained w_r. The equation then can be written as

$$w = w_i e^{-\lambda t} + w_r$$

Percolation is thus dependent on time, water content and soil properties that influence hydraulic conductivity.

The Effect of Layering on Redistribution and Percolation. The understanding of water movement through layered soils (because of their common occurrence) has been the object of a number of studies (Aylor and Parlange, 1973; Gill, 1977; Phillip, 1966; Day and Luthin, 1953; Zaslavsky, 1964; Eagleman and Jamison, 1962; Hill and Parlange, 1972; Swartzendruber, 1960).

Soil layering occurs when differences in texture, structure, clay type and bulk density exist and when land use practices result in soil surface horizons having different physical properties from subsurface ones. In the field two layering sequences are possible. The first and more common one is where a coarse-textured layer of higher hydraulic conductivity overlies a fine-textured layer of lower hydraulic conductivity.

The second and less common one is the opposite order.

In general, regarding flow through layered profiles, Miller and Gardner (1962) showed that soil matric suction must be continuous throughout the layers for water conduction. They also showed that the soil wetness and the hydraulic conductivity of layered soils may exhibit abrupt discontinuities at inter-layer boundaries. The discontinuities are found in both layering sequences although they are a result of different phenomena.

During flow in the coarse/fine profile, the initial rate of the wetting front movement is dependent only on the higher saturated hydraulic conductivity rate of the coarser layer until the wetting front reaches the finer textured layer of lower hydraulic conductivity. The lower rate of flow through the fine-textured layer is limiting, so that the portion of flow above the maximum rate of flow possible through the finer layer backs up forming a perched water table.

During downward water flow through a fine/coarse profile, water movement is initially controlled by the saturated hydraulic conductivity of the finer-textured layer until the wetting front reaches the inter-layer boundary. The wetting front movement then stops at the interlayer boundary because of differences in suction between the two layers. The greater suction in the finer-textured layer. has to be reduced to that of the coarser layer before water will flow across the interface. The suction difference is reduced by an increase in water content of the finer-textured layer. This required increase in water content is realised by the formation of a perched water table (Miller and Gardner, 1962).

Russel (1946) developed a simple mathematical description to explain the least conductive control section theory. Assuming the soil to consist of numerous homogeneous layers and that there is no surface ponding of water, the equation

$$\frac{1}{P} = \frac{1}{P_1} \times \frac{L_1}{L} + \frac{1}{P_2} \times \frac{L_2}{L} + \ldots\ldots + \frac{1}{P_n} \times \frac{L_n}{L}$$ can be derived.

According to this equation, the reciprocal of the permeability value P of any saturated column is equal to the sum of the reciprocals of the P values of all layers that comprise that column, each being multiplied by the ratio of its length to the entire length. Water flow through a layered soil profile is thus controlled by the least permeable layer (Swartzendruber, 1960).

Significance of Redistribution and Percolation in Drainage Studies. Redistribution and percolation are important in drainage studies because these processes determine the rate at which infiltrated water moves through the profile. The rates of both redistribution and percolation are also important as they determine the length of time during which upper soil horizons remain at high moisture contents after infiltration.

Horizonation is also significant in drainage because of the perched water tables that form during water flow into layered profiles. Perched water tables result in relatively higher water contents than would be the case if layering did not exist in these profiles.

EVAPOTRANSPIRATION

Introduction. Evapotranspiration is an energy requiring process whereby water is lost from the soil surface through evaporation and

transpiration. The required energy is normally supplied by solar radiation. In evaporation, water is converted from the liquid form to vapour which then escapes into the atmosphere while in transpiration, water is first taken up by plants and then lost to the atmosphere as water vapour.

Factors that Influence Evapotranspiration. Evaporation is influenced by mainly the temperature of both the air and the evaporating water, the wind velocity, the relative humidity, the nature of the evaporating surface and the quality of the evaporating water.

In addition to the factors which influence evaporation, transpiration is further influenced by the kind of plants in question, their leaf type, whether broad or narrow, succulent or not, stage of growth of the plant, rooting habits, water transporting mechanism and the thickness of the plant stand in the field (Ward, 1975).

In summary, the following factors affect evapotranspiration.

(i) The nature of the evaporative surface. Evapotranspiration is higher from free standing water than from a dry soil. The color, roughness and vegetative cover of a field will influence evapotranspiration from that field.

(ii) Wind velocity is an important factor affecting evapotranspiration since the turbulence it creates is responsible for dispersing the moist layer above the evaporating surface and thus decreasing vapour pressure and hence increasing water loss.

(iii) The relative humidity is also an important factor as it determines the vapour pressure gradient. A high relative humidity

is indicative of a low pressure gradient and thus low evapotranspiration.

(iv) Solar radiation provides the energy required to vapourise water and is therefore an important factor governing evapotranspiration. The effect of solar radiation depends on the location of the area on earth, cloud cover and atmospheric pollution. The greater the cloud cover or pollution, the less the total energy arriving at the earth's surface and so less evapotranspiration occurs.

(v) The temperature of both the air and the evaporating water is another important factor. Air temperature must be higher than water temperature for evapotranspiration to occur. The higher the air temperature relative to the water, the higher the evapotranspiration rate.

(vi) Evapotranspiration from the soil is also dependent on the soil moisture content and the hydraulic conductivity of the soil. High moisture content and a relatively higher hydraulic conductivity mean that water movement in the soil is relatively fast; thus evapotranspiration is likely to be higher in such a soil because water movement to plant roots and to the evaporating soil surface is faster than in a soil of low moisture content and relatively low hydraulic conductivity.

Evapotranspiration is important in drainage studies in that it may significantly influence water loss from the soil.

THE WATER TABLE POSITION

Introduction. The water table is the depth in the soil profile at which the soil water pore pressure is atmospheric (Tolman, 1937). The water table is the upper boundary of the zone of saturation in a profile where groundwater is unconfined (Ward, 1975). It is the plane of contact between unconfined groundwater and the capillary fringe.

A water table that is close to the soil surface may increase, through capillarity, the soil moisture content of the profile. The depth at which the water table is found and the total capillary rise will determine how much influence the water table actually has on soil moisture status. A very high water table coupled with a high capillary rise, such as may occur in high silt soils, may result in high moisture content through the entire profile.

Water table positions fluctuate with season. The water table's position is closest to the surface after a high supply of infiltration water such as might occur during spring snowmelt or high summer rainfall. During these periods the water table may actually be within the root zone creating unfavourable soil moisture conditions for plant growth.

In general, the water table is known to follow the contours of the overlying ground surface although in a subdued form (Ward, 1975; Freeze and Cherry, 1979). The amplitude of relief that the water table assumes in a given area, where precipitation infiltration is uniform over both high and low ground, will depend on the hydraulic conductivities of the soil. In coarse-textured highly conductive soils, a more horizontal surface is relatively quickly established. In fine-textured soils such

as heavy clays, whose hydraulic conductivities are low, the water table assumes a relief more closely resembling the surface relief for relatively long periods of time.

Influence of the Water Table on Drainage. The position of the water table in the profile is important in drainage studies because it may influence, through capillarity, the profile's soil moisture content and its moisture redistribution. The water table presents a drainage problem if it is within the profile or is at such a depth in the profile as to influence soil moisture in the root depth, its redistribution or percolation.

MATERIALS AND METHODS

SITE DESCRIPTION

Location. A detailed study of soil physical properties, profile water movement, variations in moisture content over time and potential evapotranspiration was undertaken at three research sites in the Falher area of the Peace River region of Alberta. The three research sites lie on a north-south line approximately 3 km west of Falher and 60 km south of Peace River. Site 1 is located 7 km north and 3 km west of Falher (SW5 79 29 W5). Site 2 is 3 km west of Falher (SE1 78 22 W5) and site 3 is 5 km west of Falher (SW25 77 22 W5). The total distance between the northernmost and southernmost sites is approximately 13 km (Figure 2).

Topography. Topography of the Falher area is very gently sloping to level. Site 1 lies in a very shallow depression although in general the land slopes very gently in a south-easterly direction. Site 2 is located on the higher part of a field which slopes to the southeast while site 3 lies on an east to west slope.

Geology. Surficial geological material in the Falher area consists of well-sorted, stratified till or modified saline lacustrine material derived from the Smoky River shales (Odynsky and Newton, 1950). The material is stone free, of grey to dark grey color. It is uniform and has numerous gypsum crystals. The deposit is generally shallow, being less than 8 m in depth.

The geological material overlying the Smokey River shales consists

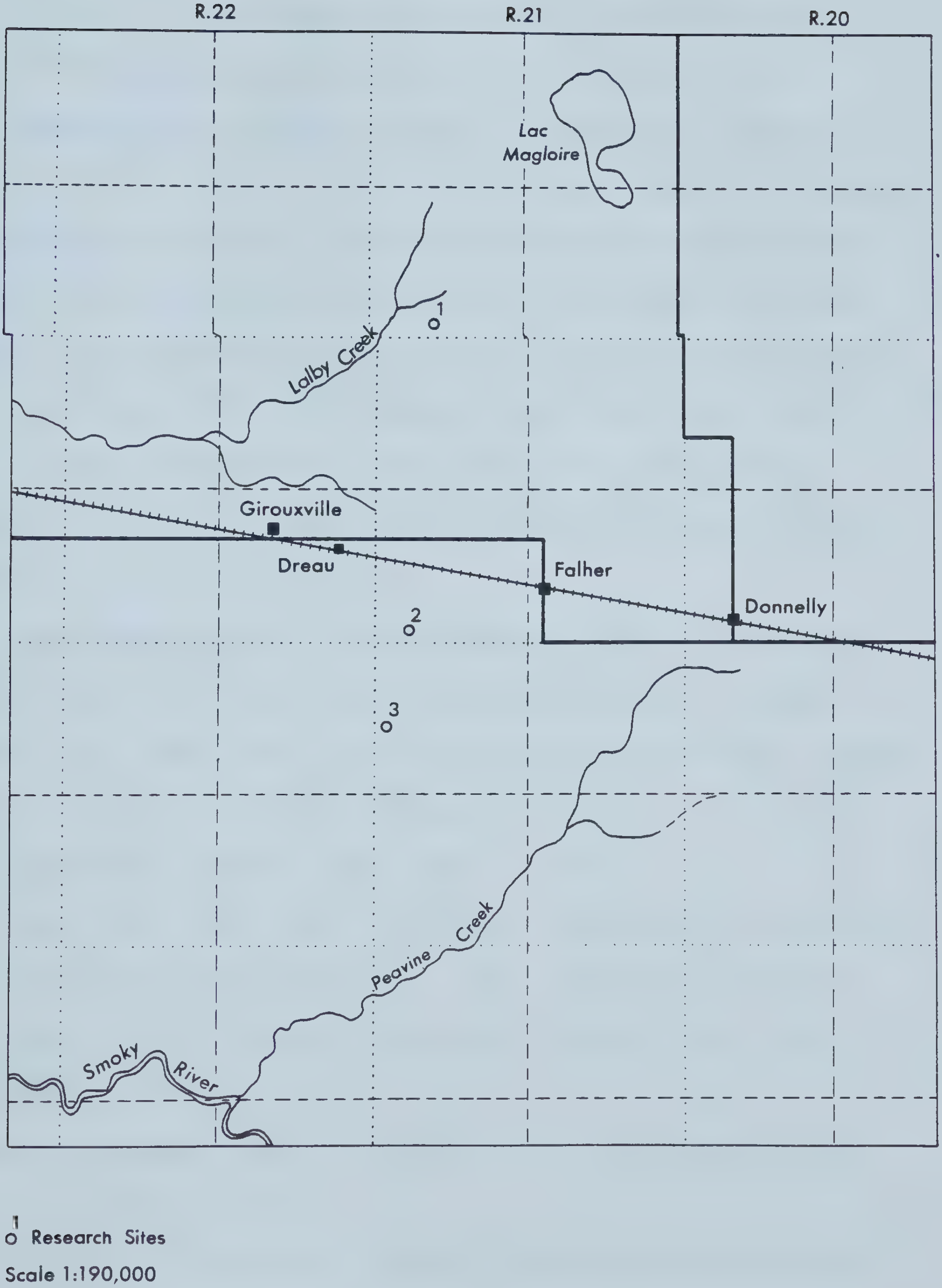

Figure 2 Location of research sites

of thin bedded dark to black shales with occasional ironstone and pyrite nodules. These shales weather readily and are thus the major constituent of the deposit.

Mineralogy of the Parent Material. Mineralogical analyses of the clay fraction of the parent material (sample depth 90 - 100 cm) are presented in Table II. Smectite/montmorillonite are the most abundant minerals followed by dioctahedral Mica (muscovite), then Kaolinite and Chlorite; the last two being almost equal in abundance.

Surface area of the clay fraction ranges from 475 to 525 m^2/g while cation exchange capacity ranges from 39 to 52 meq/100g. K content is 3% and mica constitutes an average 33% of the clay in the sample.

Water Table Position. Jones (1966) describes the possibility of groundwater in or near Falher town as unfavourable. He also reported that small quantities of water have been obtained at depths of 120 m. Most domestic and municipal water supplies in this area are from surface sources, dugout ponds and lakes.

Soils. The three sites lie within an area which has been mapped as the Falher-Rycroft association. This association is described as including Black Solodized Solonetz and Eluviated Black soils although surface horizons may be black, grey black or grey in color (Odynsky and Newton, 1950). Surface texture was reported to vary from silt loam to clay loam.

According to the current Canadian system for classifying soils (Canada Soil Survey Committee, 1978) the Falher soils would be classified as Dark Gray Solod-Solonetzic Dark Gray intergrades while the

TABLE II

MINERALOGICAL ANALYSIS OF PARENT MATERIAL

		Site 1	Site 2	Site 3
Minerals Listed in Order of Abundance		1 Smectite (Montmorillonite)	Smectite (Montmorillonite)	Smectite (Montmorillonite)
		2 Dioctahedral Mica (Muscovite)	Dioctahedral Mica (Muscovite)	Dioctahedral Mica (Muscovite)
		3 Kaolinite	Chlorite	Chlorite
		4 Chlorite	Kaolinite	Kaolinite
Mechanical Analysis	Sand %	2	2	2
	Silt %	25	21	16
	Clay %	73	77	82
Surface Area	m^2/g*	526	495	476
Cation Exch Cap	meq/100g**	39	52	46
K	%	3	3	3
Mica	%***	34	32	34

*Figures adjusted to a standard smectite sample.

**Values may be high due to dissolved ions in sample.

***Assuming mica contains 8% K and that all measured K is due to mica in the sample.

Rycroft soils would be classified as Dark Gray Solodized Solonetz - Dark Gray Solod intergrades. The soil at Site 1 has been tentatively classified as a Gleyed Solonetzic Dark Gray. The Ap and Aeg horizons are underlain by a strongly mottled Btnjg horizon at a depth of 20 cm. The soil at Site 2 has been tentatively classified as a Dark Gray Solod in which the Bnt horizon occurs at a depth of 28 cm. The soil at Site 3, tentatively classified as a Dark Gray Solodized Solonetz, is characterized by a dense Bnt horizon at a depth of 9 cm (Crown, 1982).

Cropping. All three research sites are located in cultivated fields. During the 1981 growing season, site 1 was seeded with barley, site 2 with canola and site 3 was in its fourth year of alfalfa.

SOIL PHYSICAL ANALYSIS

Semi-disturbed core samples taken for hydraulic conductivity determinations were also used to determine total porosity and bulk density. Soil samples from the same cores were used to determine particle size distribution and water retention capacity.

Six cores were taken at each site, three from the surface Ap and three from the B horizons. The cores were 7.5 cm long and 7.5 cm in diameter.

Particle Size Analysis. Particle size distribution was determined for each sample using the hydrometer method as described by Day (1965).

Soil Moisture Retention Capacity. The capacity of the soil to retain moisture, as illustrated by the soil moisture characteristic curve, was determined for each site using pressure plate apparatus.

Total Porosity. Total porosity was determined for the semi-disturbed cores using the method described by Vomocil (1965).

Sample Bulk Density. Bulk density was determined by the core method as described by Blake (1965).

Hydraulic Conductivity. Hydraulic conductivities were determined according to the falling head method described by Klute (1965) for semi-distrubed cores.

Profile Bulk Density. To determine variations in the bulk density of the soil profile, a twin probe density guage was used. The distance between the source access tube and the detector access tube was 30.5 cm. Wet density readings were taken starting at the 2.5 cm depth and progressing in 2.5 cm increments to a depth of 125 cm. Gravimetric samples were taken at the same time from similar depths to determine mass basis moisture content necessary to derive bulk density values.

INFLITRATION RATES AND WATER MOVEMENT

Infiltration Tests. Ring infiltration tests were conducted at each of the sites during early July and August, 1981, according to the double ring method described by Vertland (1965). A 5 cm head of water was allowed to drop to 3 cm before the rings were refilled. A float guage was used to monitor water levels.

Soil Water Movement. To monitor soil water movement in the profile, the twin probe density gauge was used. Two aluminum access tubes were inserted vertically into the soil at a distance of 30 cm apart. The tubes were then left in the field for at least a month to effect tight contact between the access tube and the soil. A large diameter infiltration ring was then positioned so that the two access tubes were within the ring and equidistant from the ring sides (Figure 3).

To monitor the movement of the wetting front, an initial 'standard' density reading was made prior to the ponding of water in the rings. Water was then ponded in the rings and readings were made down the tube until these coincided with the standard readings. The last depth to show a change in density was assumed to be the position of the wetting front. Readings were then repeated starting from the surface down the tube to determine changes, if any, in the profile's wet density, and to establish the new wetting front position.

The same test used to determine wetting front position was conducted to establish maximum soil moisture of the profile. In this case, water was ponded in the rings for 30 minutes before standard wet density readings were made at 5 cm intervals down the access tube to a depth of 65 cm. More readings were taken after 2, 4 and 24 hours. In this test the first set of readings, i.e., made after 30 minutes of water ponding was assumed to be the standard set. Any change in wet bulk density in the profile after this initial standard reading was attributed to a change in moisture in the profile.

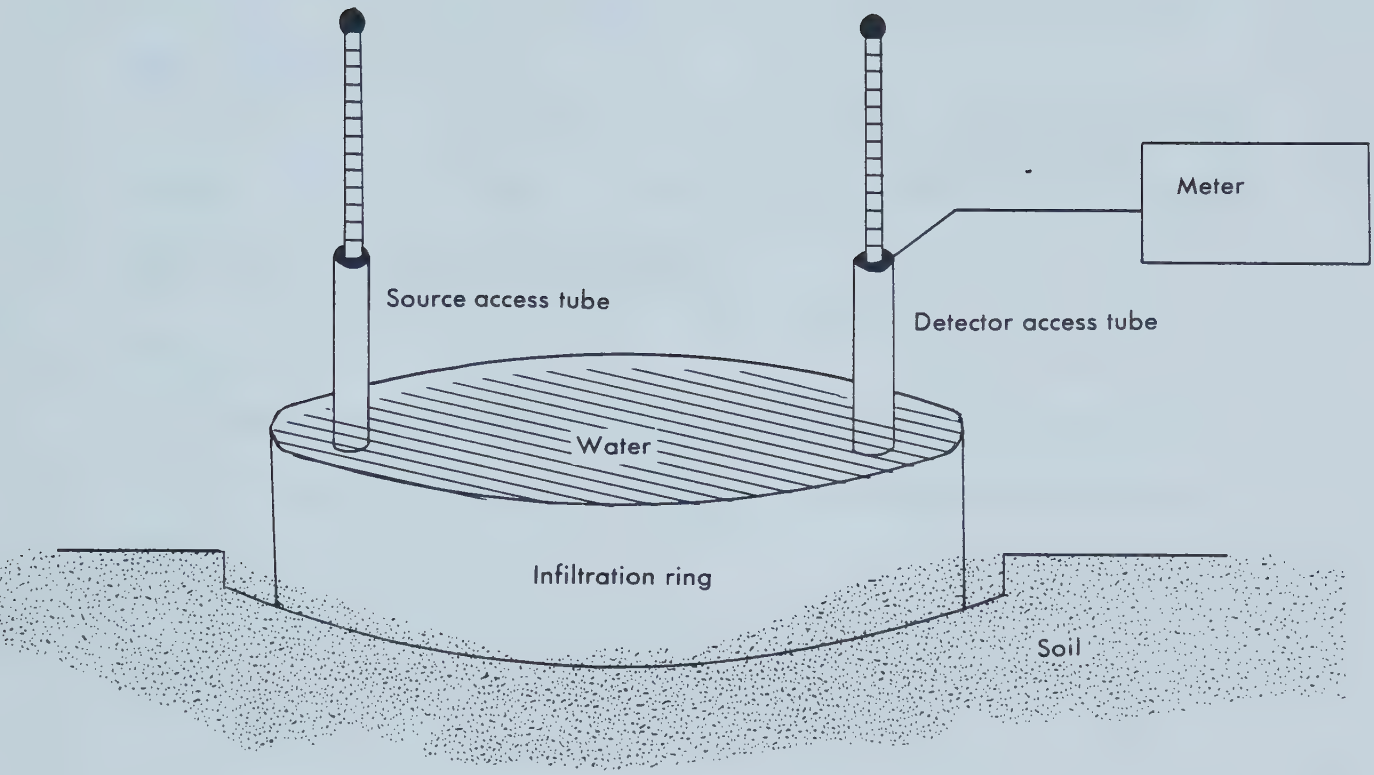

Figure 3 Field set-up of the twin probe density gauge to monitor wetting front position.

PROFILE SOIL MOISTURE
MONITORING

Profile soil moisture for the three sites for the 1981 crop season was monitored using a Campbell Pacific Nuclear neutron probe, Model 503 Depthprobe.

Readings were taken at 10 cm intervals down the tube, the first reading at the 25 cm depth and the final at a depth of 145 cm. Normally readings were made fortnightly.

RAINFALL

Rainfall measurements during the period April 28 to October 30 were collected by farmers at the site or nearby using standard rain gauges.

RESULTS AND DISCUSSION

PRECIPITATION AND EVAPOTRANSPIRATION

Precipitation and evapotranspiration have a direct influence on soil moisture and are thus important in soil moisture studies. Precipitation is a source of soil water while evapotranspiration contributes to water loss from the soil. For purposes of relating soil moisture data gathered in a single season to the long term, it is necessary to compare climatic data for the season with the long term climate data. Such a comparison for the 1981 crop season for the Falher area is given in Table III.

In general terms the area received much less total precipitation from May to October, 1981 than the long term average. Site 1 received only 40% of long term average precipitation while sites 2 and 3 received only 55 and 60% respectively of long term average precipitaiton. All the three sites received less monthly precipitation during the May to October period 1981 compared to long term mean precipitation. The only exception was site 2, which for the month of June, recorded higher precipitation than normal.

The largest deviations from long term means in precipitation in 1981 were observed during the months of August, September and October. Site 1 recorded no rainfall during this period while site 2 received 9 mm or 14% of the long term August mean, 6 mm or 16% of the long term September mean and no precipitation in October. Site 3 received 10 mm or 16% of long term precipitation

TABLE III

COMPARISONS OF 1981 PRECIPITATION
AND EVAPOTRANSPIRATION WITH LONG TERM MEANS

	Precipitation (mm)						Evapotranspiration (mm)				
	1981			Long Term			1981	Long Term			
Month	Site 1	Site 2	Site 3	Mean	Max	Min		Mean	Max	Min	% of*
May	11	22	20	38	106	01	120	101	137	81	119
June	42	79	55	63	147	20	116	122	139	99	95
July	65	46	58	68	182	26	151	129	155	115	117
August	00	09	10	63	112	21	158	102	134	079	155
September	-	06	25	38	148	12	070	051	084	026	137
October	-	00	09	27	079	00	005	011	023	002	045
Total	118	162	177	297	774	80	620	516	672	402	
% of Long Term Mean	40	56	60				120				

*% of = 1981 evapotranspiration as a % of long term mean

during August 25 mm or 65% during September and 9 mm or 32% during October.

Potential evapotranspiration for the 1981 season for Falher was 20.5% or 104 mm higher than the long term mean while May recorded 19% higher, June 5% lower, July 17% higher, August 55% higher and 37% higher and 55% lower for the months of September and October respectively.

The 1981 season was thus characterized by much lower precipitation especially in the second half of the season and much higher evapotranspiration when compared to the long term means. The soil moisture status can thus be expected to have been much lower than in a typical year.

SOIL PHYSICAL PROPERTIES

Particle Size Distribution. Despite variations in actual clay, silt and sand quantities between the three sites, the soils at these sites fall in the clay and heavy clay textural classes (Canada Soil Survey Committee, 1978) (Table IV).

Clay contents in these soils are high, at least 50%, while the sand content are generally low, below 10%. Exceptions are samples of the Ap horizon at site 2 and a B horizon sample from site 1. Similar results have been reported for the parent material by Chanasyk *et al*. (1981).

The Ap samples have less clay and relatively higher sand contents than do the B horizon samples. This is true for all samples analysed except sample B-1 (a B horizon sample from site

TABLE IV

PARTICLE SIZE DISTRIBUTION

OF FALHER SOIL SAMPLES

Site	Horizon & Sample #	Particle Size Distribution Sand %	Silt %	Clay %	Texture
1	Ap - 1	7	32	61	Heavy Clay
1	Ap - 2	7	31	62	Heavy Clay
1	Ap - 3	7	32	61	Heavy Clay
1	B - 1	12	31	57	Clay
1	B - 2	2	10	88	Heavy Clay
1	B - 3	2	13	85	Heavy Clay
2	Ap - 1	10	36	54	Clay
2	Ap - 2	12	35	53	Clay
2	Ap - 3	11	38	51	Clay
2	B - 1	7	30	63	Heavy Clay
2	B - 2	3	31	66	Heavy Clay
2	B - 3	6	26	68	Heavy Clay
3	Ap - 1	11	23	66	Heavy Clay
3	Ap - 2	10	23	67	Heavy Clay
3	Ap - 3	9	23	68	Heavy Clay
3	B - 1	3	17	80	Heavy Clay
3	B - 2	5	15	80	Heavy Clay
3	B - 3	11	18	71	Heavy Clay

1) which had more sand and relatively less clay than the Ap horizon samples.

Actual clay content differences between the Ap horizon and B horizon samples varied from site to site. This clay content difference averages 15% at site 1, 13% at site 2 and 10% at site 3.

Sand contents were higher in the Ap horizon samples when compared to the B horizon samples. The lowest sand content determined is 2% in two of the B horizon samples of site 1 while the highest, 12% is found in samples B-1, a B horizon site 1 sample, and Ap-2, an Ap horizon sample from site 2.

Silt contents are higher in the Ap horizon samples compared to B horizon samples. The range for silt contents is from 10% in sample B--2 from site 1 to 38% for sample Ap-3 from site 2.

Moisture Retention Capacity. The moisture retention capacity of the soil, usually presented as the moisture characteristic curve, is a soil physical property that is well correlated to its particle size distribution. Generally, the higher the clay content, the higher the water retention capacity of the soil (Hillel, 1980a; Schwab *et al*., 1966). The smaller sized clay particles result in higher total porosity and a relatively higher proportion of this porosity being taken up by hygroscopic moisture.

Table V gives results of the water retention capacity tests. Water retention curves for Ap and B horizon samples for sites 1,2 and 3 are presented as Figures 4, 5 and 6 respectively. Marked

TABLE V

SOIL WATER RETENTION CAPACITIES
FOR THE A AND B HORIZONS OF FALHER SOILS

Site	Horizon & Sample No.	% Soil Moisture Retained: Pressure (kPa x 10^2) 0.33	1	2	3	6	9	12	15
1	Ap - 1	31	28	27	23	21	24	20	23
1	Ap - 2	30	28	26	23	20	20	18	20
1	Ap - 3	31	28	25	23	20	21	-	21
1	B - 1	23	21	17	18	16	15	15	13
1	B - 2	37	32	30	29	27	25	25	25
1	B - 3	37	32	21	28	25	25	27	23
2	Ap - 1	30	26	24	20	18	21	15	17
2	Ap - 2	29	24	22	22	18	21	20	18
2	Ap - 3	29	26	24	21	19	18	17	17
2	B - 1	29	25	24	22	20	23	17	20
2	B - 2	32	29	28	26	23	24	23	23
2	B - 3	33	29	27	25	21	23	15	22
3	Ap - 1	29	26	24	24	20	19	19	19
3	Ap - 2	29	27	25	23	20	24	18	19
3	Ap - 3	29	25	24	22	20	19	19	18
3	B - 1	30	29	28	23	21	23	16	20
3	B - 2	31	28	26	24	22	21	21	21
3	B - 3	27	24	23	21	19	21	17	20

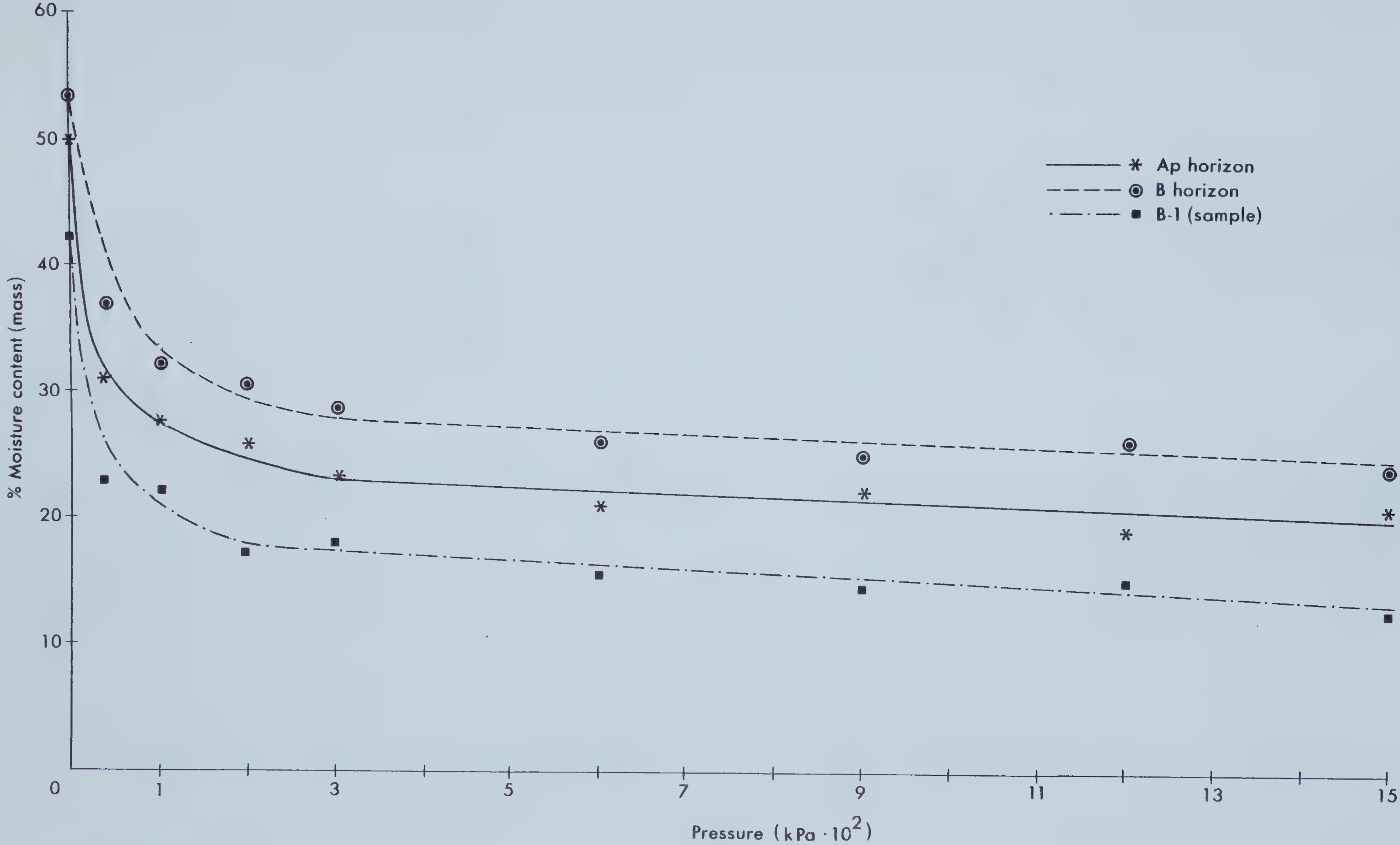

Figure 4 Soil moisture retention curves Site 1

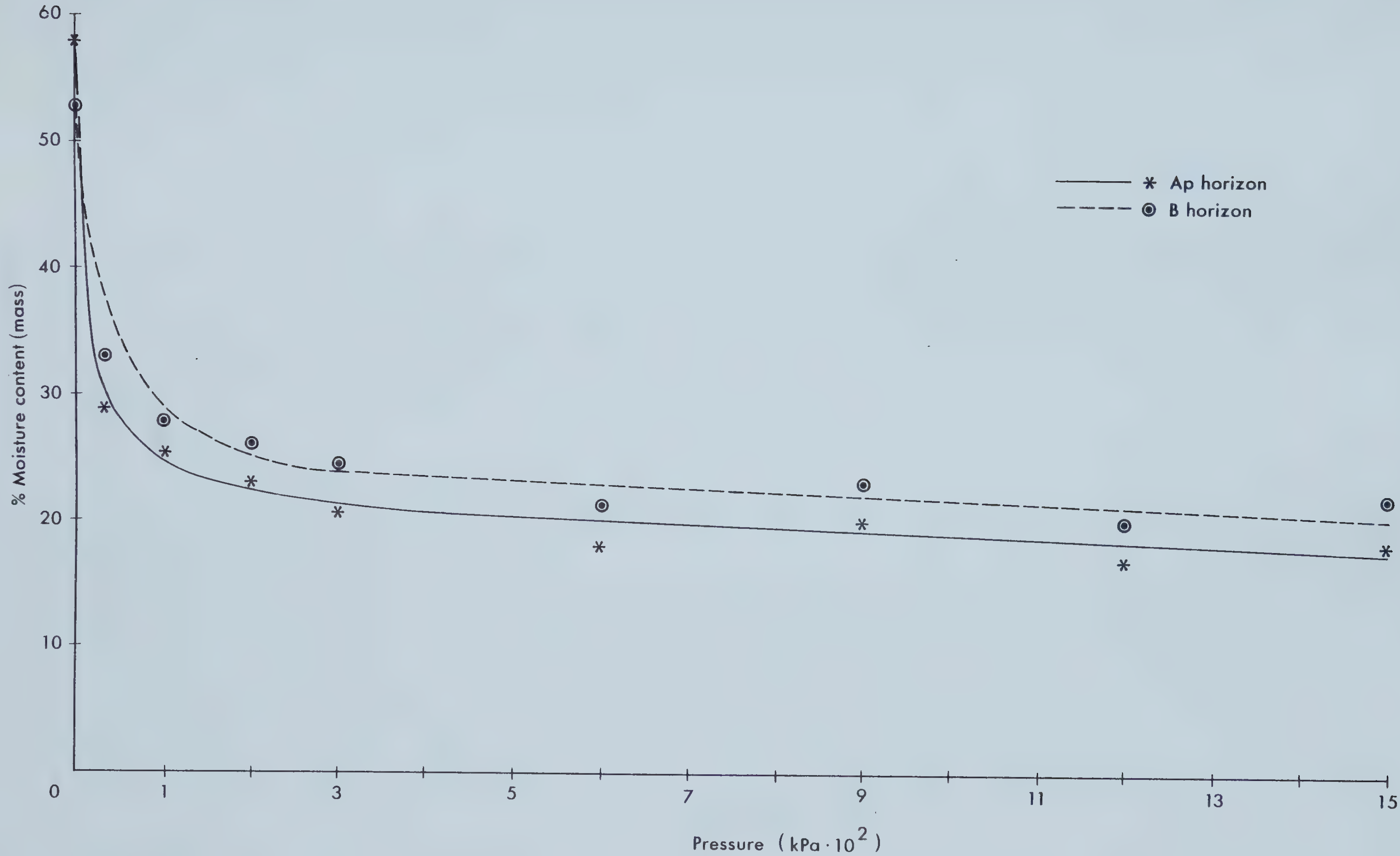

Figure 5 Soil moisture retention curves Site 2

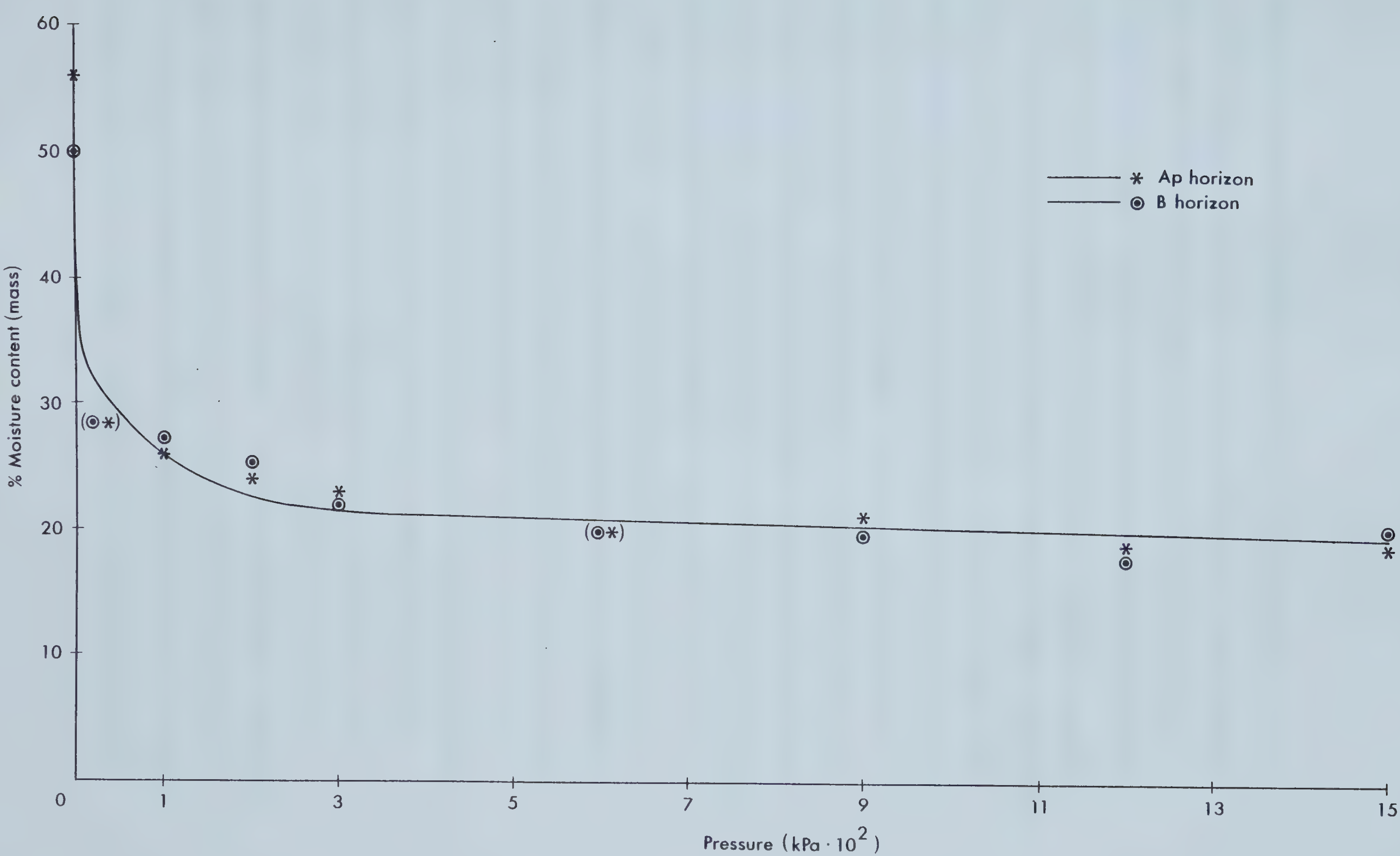

Figure 6 Soil moisture retention curves Site 3

differences between Ap and B horizon samples are observed at sites 1 and 2 (Figures 4 and 5) but not site 3(Figure 6). Note however that sample B-1 from site 1 gives a much lower retention curve. At both sites 1 and 2 with the exception of the curve B-1 at site 1, the lower lying B horizon has a higher water retention capacity.

The observed differences in moisture retention capacities between the Ap and the B horizon samples may be explained on the basis of clay content. Average clay content differences between the two horizons are 25% at site 1 if sample B-1 is excluded and 15% if included; 13% at site 2 and 10% at site 3. The smaller difference in clay content between the Ap and B horizon samples at site 3 may be one reason for the lack of a clear distinction between the moisture retention curves of the two horizons at this site.

The lack of a clear distinction between the Ap and the B horizon curve for site 3, (Figure 6) may also be due to different field management practices, such as depth of tillage which may have caused the mixing of the two horizons.

The moisture retention curve also gives an indication of pore size distribution in a given soil (Hillel, 1980a). Soils whose moisture retention curve slopes are low, i.e. flat curves, have a more uniform pore size distribution. The curves in Figures 4, 5 and 6 are relatively flat between 300 and 1500 kPa. These soils therefore are likely to have uniform pore size distribution.

High moisture retention at 1500 kPa (20-25% in site 1, 15-22% in site 2 and 16-21% in site 3 samples) is an indication of soils

whose pore geometry is predominated by finer micropores.

Porosity. Water movement through a given soil is influenced by total porosity and pore size distribution. The rate of water movement is greater in soils whose pore geometry is dominated by macro- rather than micropores (Hillel, 1980a; Schwab *et al.*, 1966).

Unlike particle size distribution and water retention capacity results, no general trends are observable in the total porosity values (Table VI). Total porosity is not a permanent soil property as it is modified by changes in soil structure and thus by soil management practices. With the exception of sample B-1 from site 1 whose total porosity is 42%, the samples have total porosities of 50% or greater. Site 1 results show no difference in total porosities between the Ap and B horizon when sample averages are used for comparison. However, if sample B-1 is excluded, the B horizon has higher total porosity. At site 2 and 3 the opposite is true, the Ap horizons show higher total porosities.

Aeration porosities (the proportion of total porosity which is air filled when moisture content is at field capacity) for Falher soils estimated from total porosity figures (Table VI) and field capacities (Figures 4,5 and 6) are about 22% for the Ap and 13% for the B horizon at site 1, 29% for the Ap horizon and 21% for the B horizon at site 2, and 27% for the Ap and 21% for the B horizon at site 3. At all the three sites, aeration porosities of the B horizon samples are lower than those of the apparent Ap horizon samples. Thus where adequate amounts of water to saturate fully the soil are available at the soil surface, perched water tables

TABLE VI

SELECT SOIL PHYSICAL PROPERTIES

Site	Horizon & Sample No.	Total Porosity %	Bulk Density g/cm^3	Hydraulic Conductivity cm/sec
1	Ap - 1	52	1.26	2.23×10^{-3}
1	Ap - 2	53	1.33	5.06×10^{-7}
1	Ap - 3	53	1.28	9.92×10^{-7}
1	B - 1	42	1.72	4.21×10^{-8}
1	B - 2	62	1.42	5.47×10^{-8}
1	B - 3	56	1.38	3.18×10^{-8}
2	Ap - 1	58	1.20	7.31×10^{-4}
2	Ap - 2	59	1.12	5.18×10^{-3}
2	Ap - 3	57	1.03	-
2	B - 1	52	1.31	6.23×10^{-8}
2	B - 2	54	1.34	5.21×10^{-8}
2	B - 3	53	1.46	2.29×10^{-8}
3	Ap - 1	53	1.46	7.53×10^{-8}
3	Ap - 2	55	1.50	5.79×10^{-8}
3	Ap - 3	61	1.39	6.94×10^{-7}
3	B - 1	50	1.53	3.28×10^{-8}
3	B - 2	52	1.42	3.63×10^{-8}
3	B - 3	49	1.53	2.30×10^{-8}

maydevelop in the Ap horizon because of the likely lower flow rates of the B horizons.

The higher aeration porosities in the Ap horizons may be due to the higher sand content and thus proportionally greater percentage of macropores.

Bulk Density. Bulk dẹnsity, which is the ratio of the mass to the bulk volume of the soil particles plus pore spaces in the soil, like total porosity, is not an invariant characteristic for a given soil. It varies with soil structure and hence with soil management practices (Blake, 1965).

Bulk density values determined on semi-disturbed cores show that samples from the Ap horizons have lower bulk densities than those from B horizons (Table VI). Similar results were obtained using the twin probe density guage (Figure 7).

Profile bulk densities are lowest in the Ap horizons with a sharp increase to a maximum of 1.61 g/cm^3 at a depth of 12.5 cm at site 1, 1.51 g/cm^3 at a depth of 17 cm at site 2 and 1.61 g/cm^3 at a depth of 24 cm at site 3. Below this maximum bulk density point, the profile bulk density decreases slightly and except for site 2, assumes a more or less constant value throughout the remainder of the profile.

The site 2 bulk density profile shows a marked decrease in bulk density at the 60 cm depth of 0.2 g/cm^3. The higher bulk densities of the B horizons are consistent with the particle size analysis and aeration porosity data (presented on page 44).

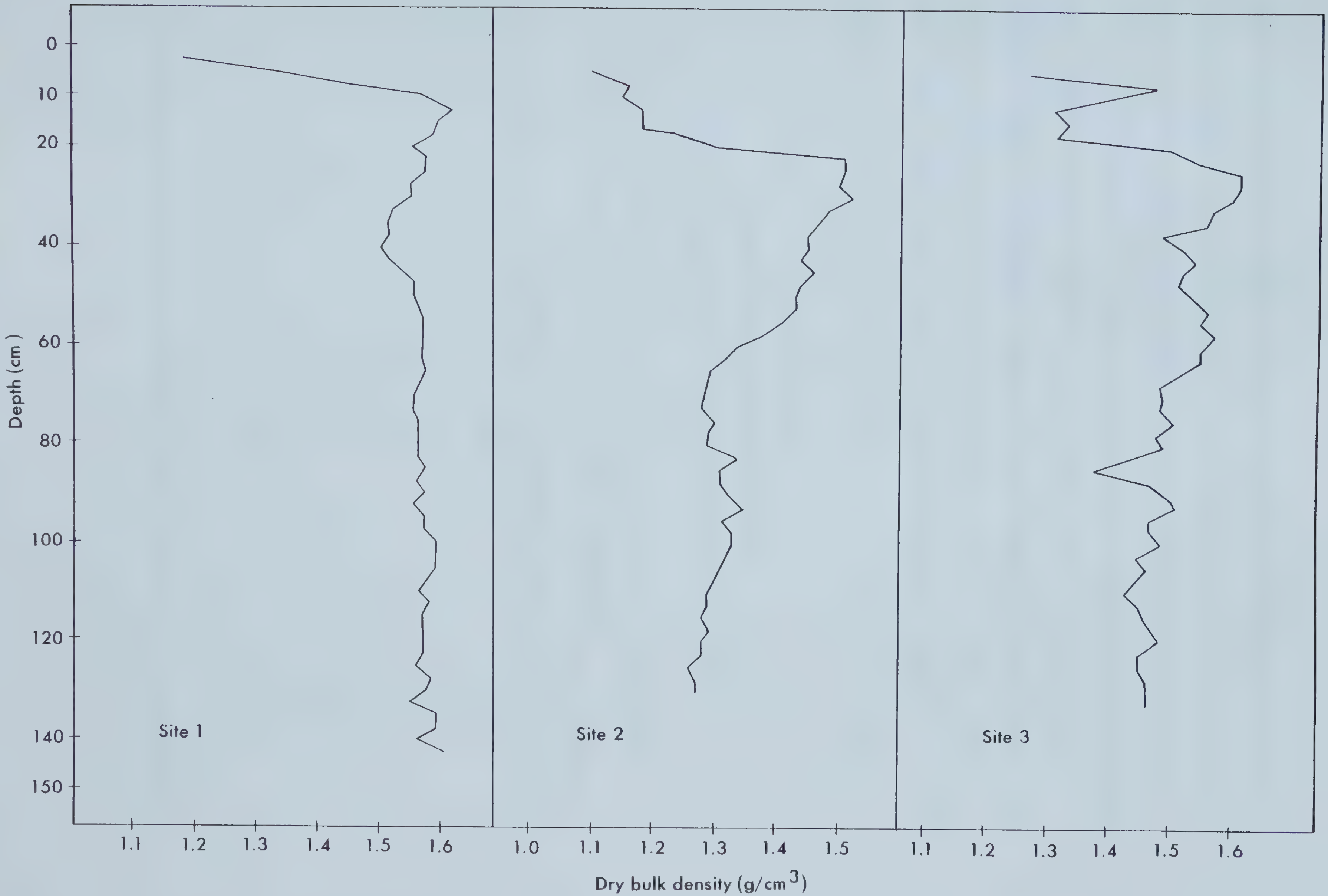

Figure 7 Profile Bulk densities at the three sites

Relatively high clay contents coupled with low aeration porosities indicate high bulk densities while high sand contents together with high aeration porosities indicate relatively lower bulk densities.

In a comparison of bulk densities determined by the core method and by the twin probe density guage (Table VII), the twin probe density guage gave values between 4 and 9% higher. Only in the case of the site 3 Ap horizon did the semi-distrubed sample show a 4% higher bulk density than the twin probe density guage.

TABLE VII

BULK DENSITY DETERMINATIONS

SEMI-DISTURBED CORES vs TWIN PROBE DENSITY GUAGE

Site	Horizon	Bulk Density in g/cm^3	
		Mean of Core Samples	Mean Twin Probe Guage
1	Ap	1.29	1.42
1	B	1.51	1.57
2	Ap	1.12	1.18
2	B	1.37	1.50
3	Ap	1.45	1.39
3	B	1.49	1.58

Hydraulic Conductivity. Hydraulic conductivity is perhaps the most important soil property involved in the flow of water (Klute, 1965). Saturated hydraulic conductivity is an indication of the potential maximum rate of water flow through the soil.

Hydraulic conductivities of the soils studied are extremely low (Table V). Hydraulic conductivities of the Ap horizon samples at site 1 range from 2×10^{-3} cm/sec to 1×10^{-8} cm/sec, while B horizon samples are less variable with a mean of 4×10^{-8} cm/sec. Results for site 2 show a similar trend with Ap horizon samples having hydraulic conductivities 5×10^{-3} cm/sec and 7×10^{-4} cm/sec while the B horizon samples all show hydraulic conductivities of 10^{-8} cm/sec. Results for site 3 show little difference between the Ap and B horizon samples.

The trends shown for sites 1 and 2 can be explained by the particle size analysis, porosity and bulk density data. Lower bulk densities, relatively higher aeration porosities and lower water retention capacities in the Ap horizon favour relatively higher rates of water flow. Site 3 results may partly be attributed to the lower bulk density and minor water retention capacity differences between the Ap and B horizon samples. It is also possible that at site 3 the minor differences between the Ap and B horizon result from management practices. This field has been under alfalfa for several years and the structure of the Ap horizon may have resulted from a lack of annual cultivation and/or vehicle movement over the field during haying operations.

A third possible explanation for the lack of difference between the Ap and B horizon hydraulic conductivities may be that because of the relatively shallow depth to the B horizon (9 cm), when the field is cultivated more B horizon material is mixed with the surface horizon than at the other sites. Thus the Ap horizon has properties that are similar to the B horizon.

In general, hydraulic conductivities of 10^{-7} to 10^{-8} cm/sec suggest extremely slow rates of water movement, so that water flow through the soils at all sites is practically nil.

INFILTRATION RATES

Final infiltration rates recorded vary from a low 4 mm/h at site 1 to a high 30mm/h at site 3 with the final infiltration rate at site 2 being 22 mm/h (Table VIII).

These results would appear, with the exception of site 1, to indicate that infiltration is not the major reason for the drainage problem. This observation may not be entirely correct because of the soil conditions at the time of the tests. The soils at all the sites were dry and cracked at the time of these tests and substantial lateral flow was noted at the A-B horizon interface through these cracks once water was ponded on the surface. This appears to be the reason for the high infiltration rates determined.

In an attempt to reduce lateral flow, test rings were driven deeper into the soil during the second set of infiltration tests (test 2) and water was artificially added to the soil the day before the tests were run to reduce cracking. The results of this

TABLE VIII

INFILTRATION RATES
AT THE 3 FALHER SITES

Site	Test	Initial Surface Moisture %	Infiltration in mm/h						
			t = 1	t = 10	t = 30	t = 60	t = 90	t = 120	t = 220
	1	13	420	96	48	24	8	8	4
1	2	39	360	108	48	8	6	6	6
	*3	31	0	0	0	0	0	0	0
	1	16	300	132	78	64	60	52	36
2	2	32	120	60	36	24	24	24	22
	*3	33	180	0	0	0	0	0	0
	1	14	240	96	96	84	75	75	72
3	2	33	360	180	108	72	36	36	30
	*3	29	420	0	0	0	0	0	0

*Test conducted after removal of some of the A horizon.
t is time in minutes after onset of infiltration

second set of tests show much lower infiltration rates at sites 2 and 3 (Table VIII).

It was also noted from the site 1 tests that lower infiltration rates were recorded when infiltration rings were driven into the B horizon of the soil. To confirm the observation, 7 to 8 cm of the Ap soil material was removed to ensure that the rings were driven into the B horizon at each of the sites. Test results under these conditions (shown as test 3 in Table VIII) show that within a few minutes the infiltration rate had dropped to zero.

These results show that infiltration into the apparent Ap soil horizon occurs at a rapid rate especially when the soil is initially dry. Infiltration into the B horizon, however, drops to zero. Soil cracking greatly increases initial infiltration rates. Thus infiltration into these soils occurs at a rapid initial rate, especially if the soil is dry and cracked, until the wetting front reaches the A-B horizon interface, after which infiltration slows down and finally ceases.

WATER MOVEMENT THROUGH THE PROFILE

Results of the tests to monitor water movement through the profile during infiltration using the twin probe density guage are presented in Figures 8 and 9. Figure 8 results are for the test to monitor the wetting front's position. Figure 9 shows the longer 24 hour test results aimed at establishing soil moisture content after longer periods of time under ponding.

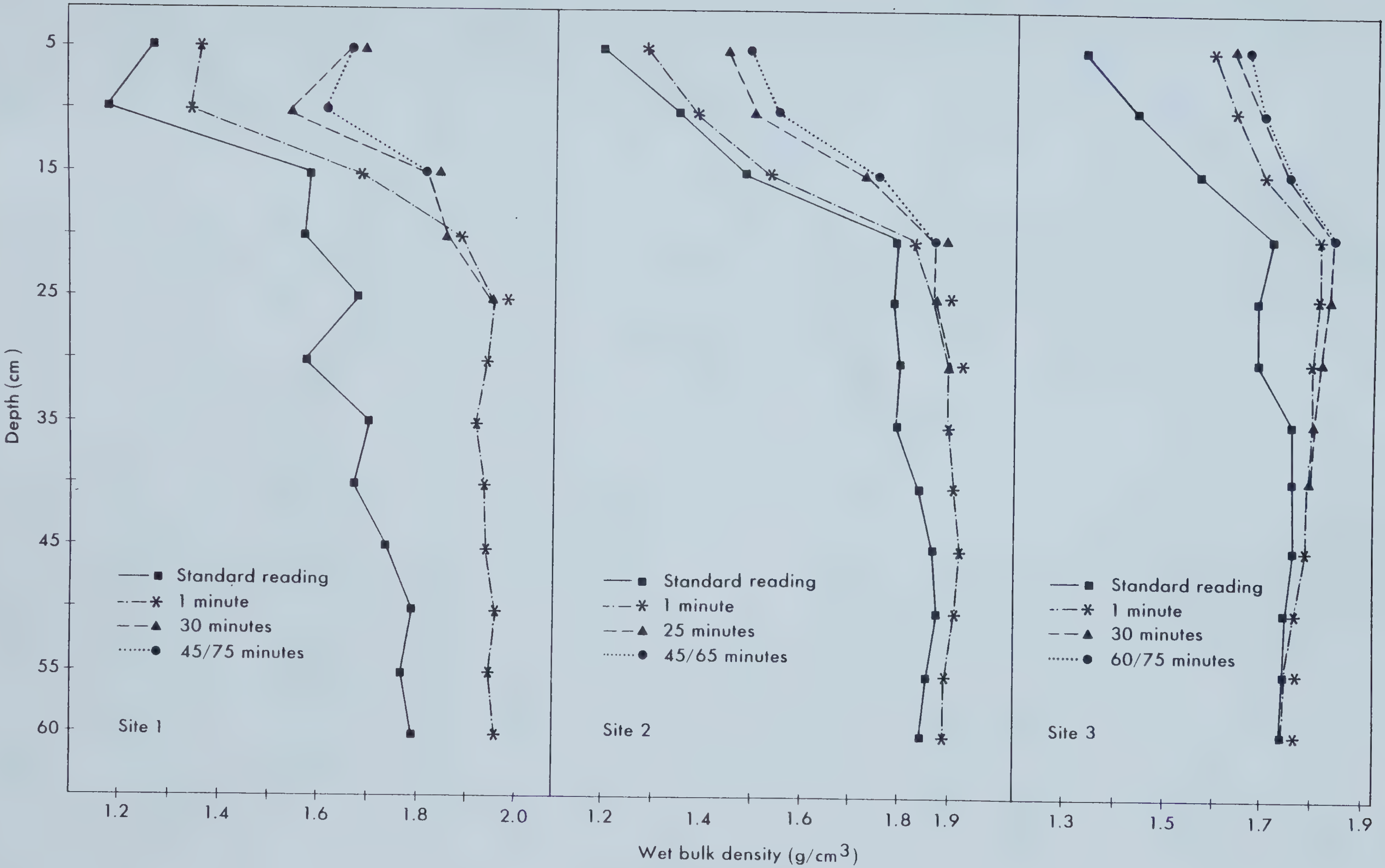

Figure 8 Water movement in the 3 soil profiles

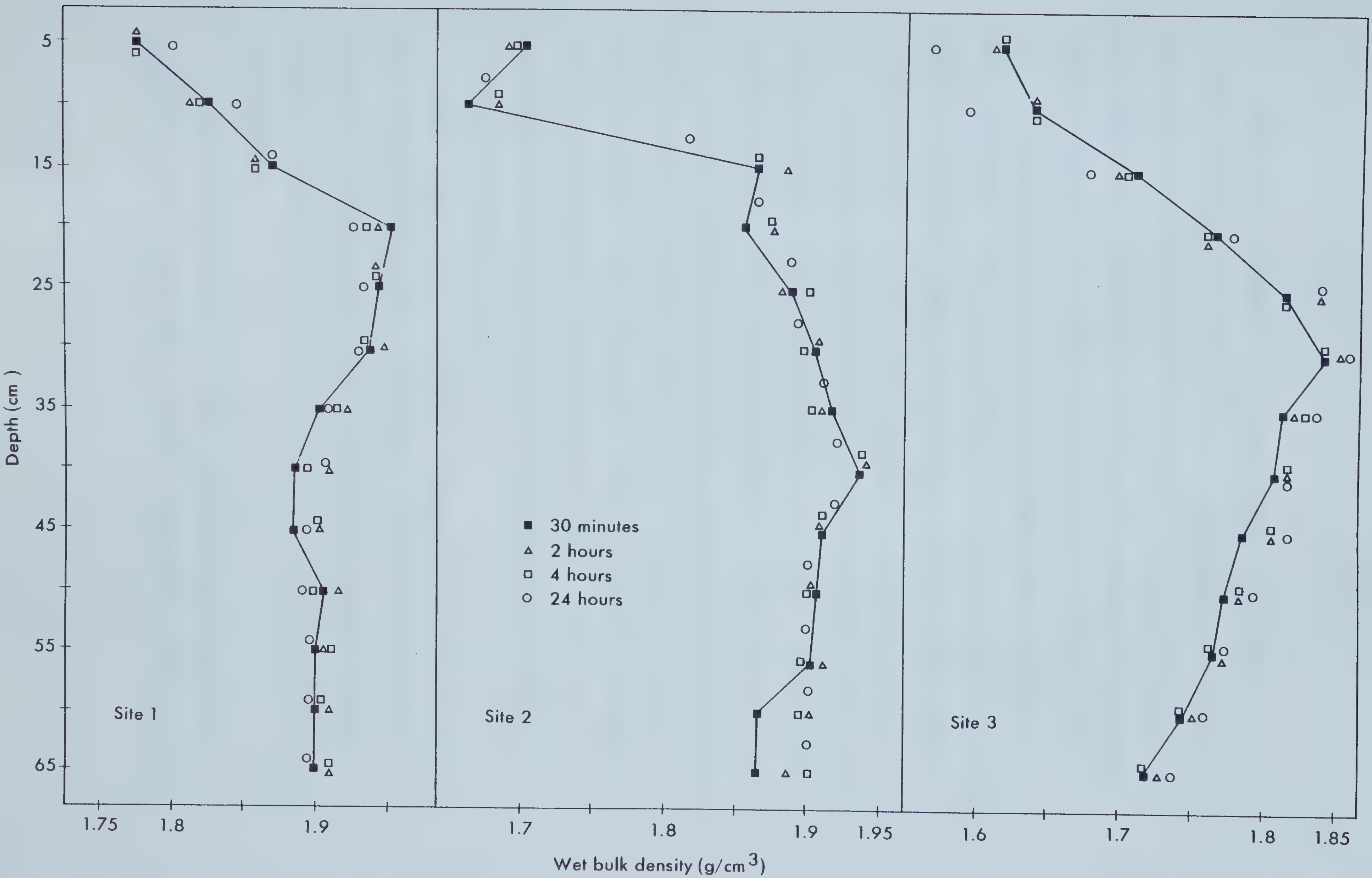

Figure 9 Water distribution over a 24 hour period

There was an increase in wet bulk density and hence moisture content of the profile within the first few minutes after water was ponded at the soil surface (Figure 8). It is unlikely, however, that this initial increase in soil moisture represents the wetting front's position. The increase in soil moisture is more likely due to water moving down the profile through cracks in the soil and along access tube sides. After this initial increase in profile moisture, there was a continued increase in the moisture content in the top 20-25 cm until a final moisture content was reached.

These tests show water moved to varying depths in these profiles, at least 60 cm in sites 1 and 2 and 50 cm in site 3. The tests further show that after the initial increase in soil moisture there were further increases of moisture with time in the surface horizon. These subsequent increases which were restricted to the apparent Ap horizon were a result of water backing up, and they continued until a final moisture content had been reached. Thereafter very little water entered the soil and further additions of water led to increased surface ponding.

From a comparison of the data in Figures 8 and 9 the final wet bulk densities and hence moisture contents in the profiles are similar. This suggests the profiles attained their maximum moisture contents within the first few hours after ponding water on the surface.

The observations made in these redistribution tests may be explained on the basis of soil physical properties. The high water

retention capacity of these soils effectively reduces the proportion of soil stored water that may be depleted naturally by evapotranspiration while the extremely low hydraulic conductivity rates, especially in the B horizons, practically eliminate drainage of water through the soil. As a consequence, the proportion of total porosity that may be air filled at a given time is reduced, thereby reducing the total amount of water required to saturate the soil. Once saturated, the soil will remain so until moisture is reduced through evapotranspiration as hydraulic conductivities are low. Additions of water to the soil while it is in this saturated state leads to surface ponding because of the negligible rate of deep percolation.

This also explains why the final infiltration rates in these soils are so low.

SEASONAL PROFILE MOISTURE

Soil moisture contents during the 1981 season were different between the three sites (Figure 10). Moisture was highest in all the three soils following snowmelt in late April and declined sharply during most of the month of May. Between late May and early August no significant changes were recorded until late August when total soil moisture decreased sharply so that by mid-September all profiles recorded their lowest moisture contents fof the season. Total soil moisture showed a slight increase at the end of October.

Sites 1 and 2 were at higher moisture contents than site 3 throughout the season. Soil moisture contents at sites 1 and 2

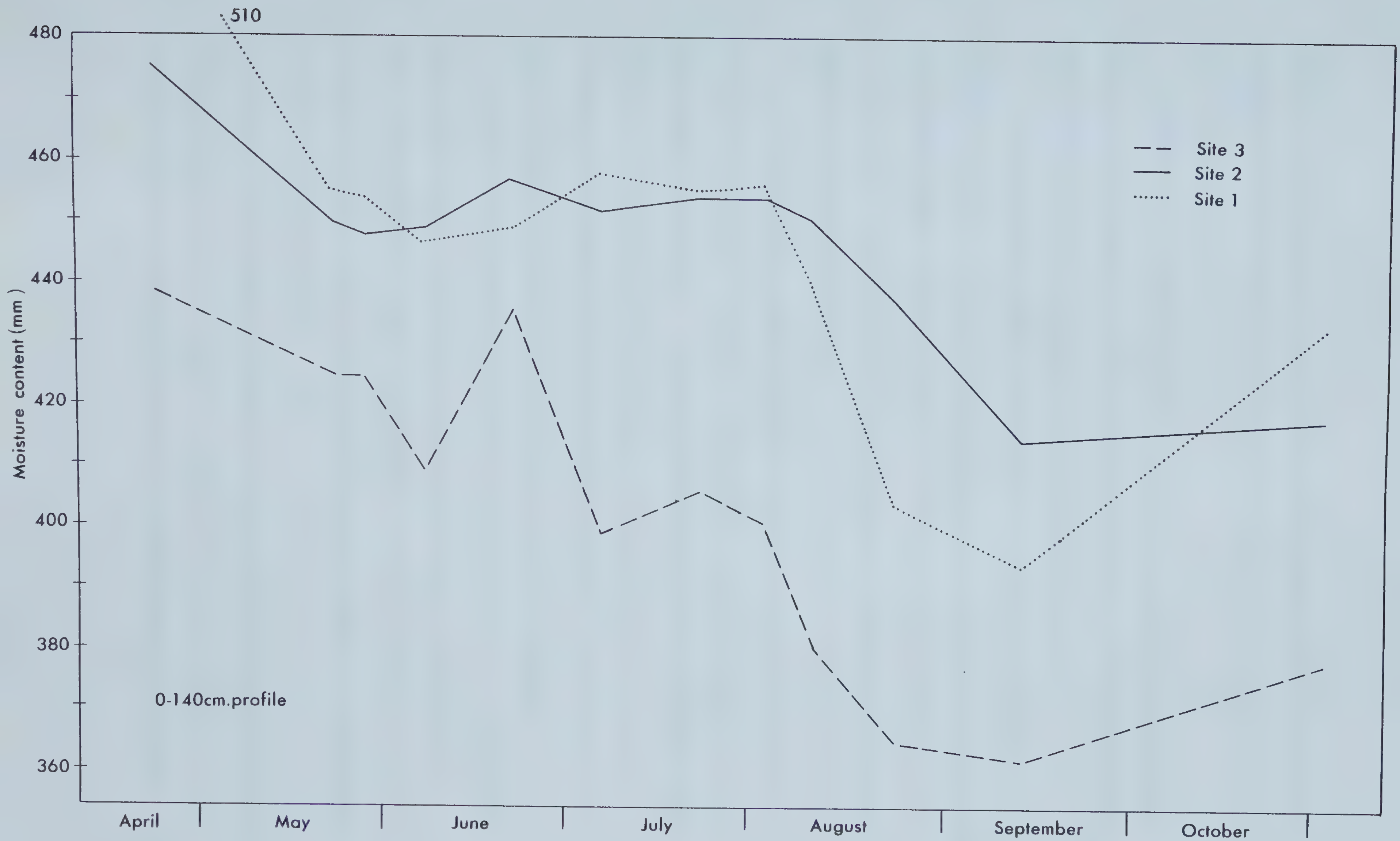

Figure 10 Total profile soil moisture at Falher 1981

were similar during the early part of the season until the August-September period when site 1 registered much less moisture than did site 2.

From the bi-weekly profile moisture readings it was observed that moisture contents of the B and C horizons of the soils showed much less variations compared to the surface layers.

To determine to what depth down the profile variations in moisture content occurred, the total profile was arbitrarily divided into three layers; layer 1 being 0 - 50 cm, layer 2 being 50 - 100 cm and layer 3 being 100 - 140 cm. Moisture variations in these three layers were considered.

Of the moisture changes in the three layers for each of the soils, layer 3 shows the least variation in moisture content during the season (Figure 11). The maximum change in moisture from that recorded at the start of the season is 22 mm or 6% for site 1, 11 mm or 3% for site 2 and 7 mm or 2% for site 3. Thus there was little change in moisture content below the 100 cm depth.

Variations in moisture content in the 50 - 100 cm layer shows a maximum change in moisture content of 21 mm or 4% at site 1, 6 mm or 1% at site 2 and 14 mm or 3% at site 3. Moisture content of the 50 - 100 cm layer, therefore did not change very much during the 1981 cropping season.

The 0 - 50 cm layer showed the most change in moisture content. The maximum decrease in moisture content shown by the site 1 curve is 75 mm or 15% while site 2 shows a maximum change in moisture

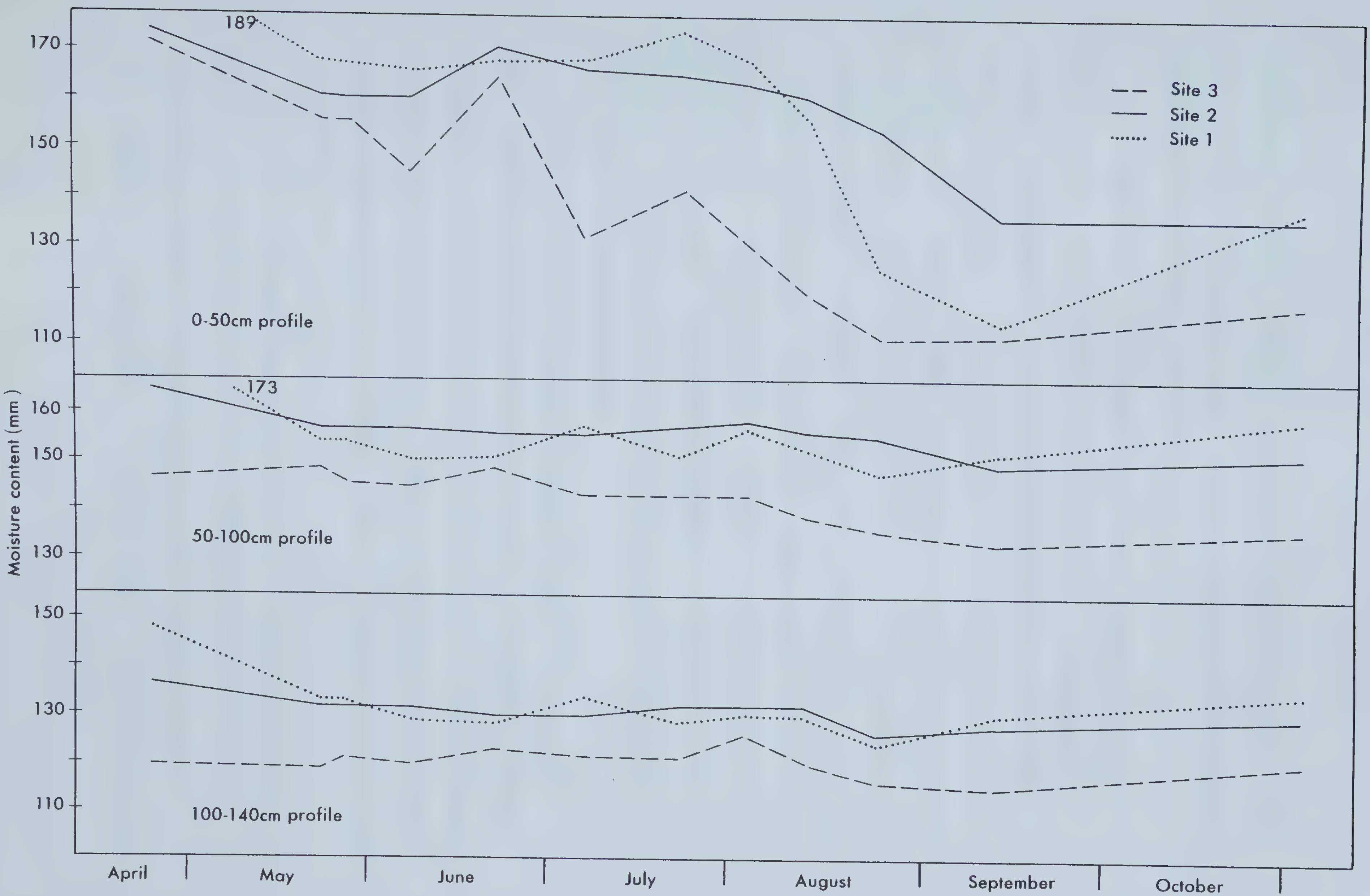

Figure 11 Profile soil moistures of three layers at the 3 Falher sites

of 35 mm or 7% and site 3,60 mm or 12%. These observations suggest that most of the changes in soil moisture content during the 1981 season occurred within the top 50 cm.

A comparison of the changes in moisture content in the 0 - 30 cm and 30 - 50 cm layers at the three sites is presented in Table IX and Figure 12. The maximum change for the 1981 season was 52 mm or 17% at site 1, 28 mm or 9.5% at site 2 and 42 mm or 14% at site 3, while it was 24 mm or 12% at site 1, 100 mm or 5% at site 2 and 19 mm or 9.5% at site 3 in the 30 - 50 cm layer. The contribution of the 30 - 50 cm layer to maximum moisture change in the 0 - 50 cm layer during the 1981 season was 31% at site 1, 27% at site 2 and 30% at site 3. Changes in moisture content in the 30 - 50 cm layer are therefore substantial and contribute significantly to moisture changes in the 0 - 50 cm layer and to total profile moisture changes.

Trends in moisture content change in the 0 - 30 cm and the 30 - 50 cm layer are illustrated in Figure 12. Moisture content change in the 30 - 50 cm layer is similar but much less pronounced compared to the 0 - 30 cm layer. Also illustrated is the delayed response to soil moisture changes in the 0 - 30 cm layer that occurs in the 30 - 50 cm layer.

The lack of significant change in total moisture content in the 50 - 100 cm and the 100 - 140 cm layer of the three profiles is due to low hydraulic conductivities which restrict downward water drainage in these soils. The predominantly fine pore geometry

TABLE IX

MOISTURE CONTENT IN THE 0-30 cm & 30-50 cm
LAYERS FOR THE PERIOD 27 APRIL TO 2 NOVEMBER 1981

		Site 1		Site 2		Site 3	
		Moisture Content in: 0-30 cm	30-50 cm	Moisture Content in: 0-30 cm	30-50 cm	Moisture Content in: 0-30 cm	30-50 cm
Month	Day	mm	mm	mm	mm	mm	mm
04	27	111	78	105	68	102	70
05	24	98	72	93	69	89	68
05	28	96	72	93	68	89	67
06	08	100	67	93	67	80	64
06	22	101	68	102	69	95	69
07	07	97	72	98	69	73	59
07	23	103	70	98	68	81	60
08	03	96	71	96	69	72	68
08	11	91	66	93	68	61	55
08	25	71	56	86	67	60	51
09	14	59	54	78	58	61	51
11	02	77	61	77	59	66	54

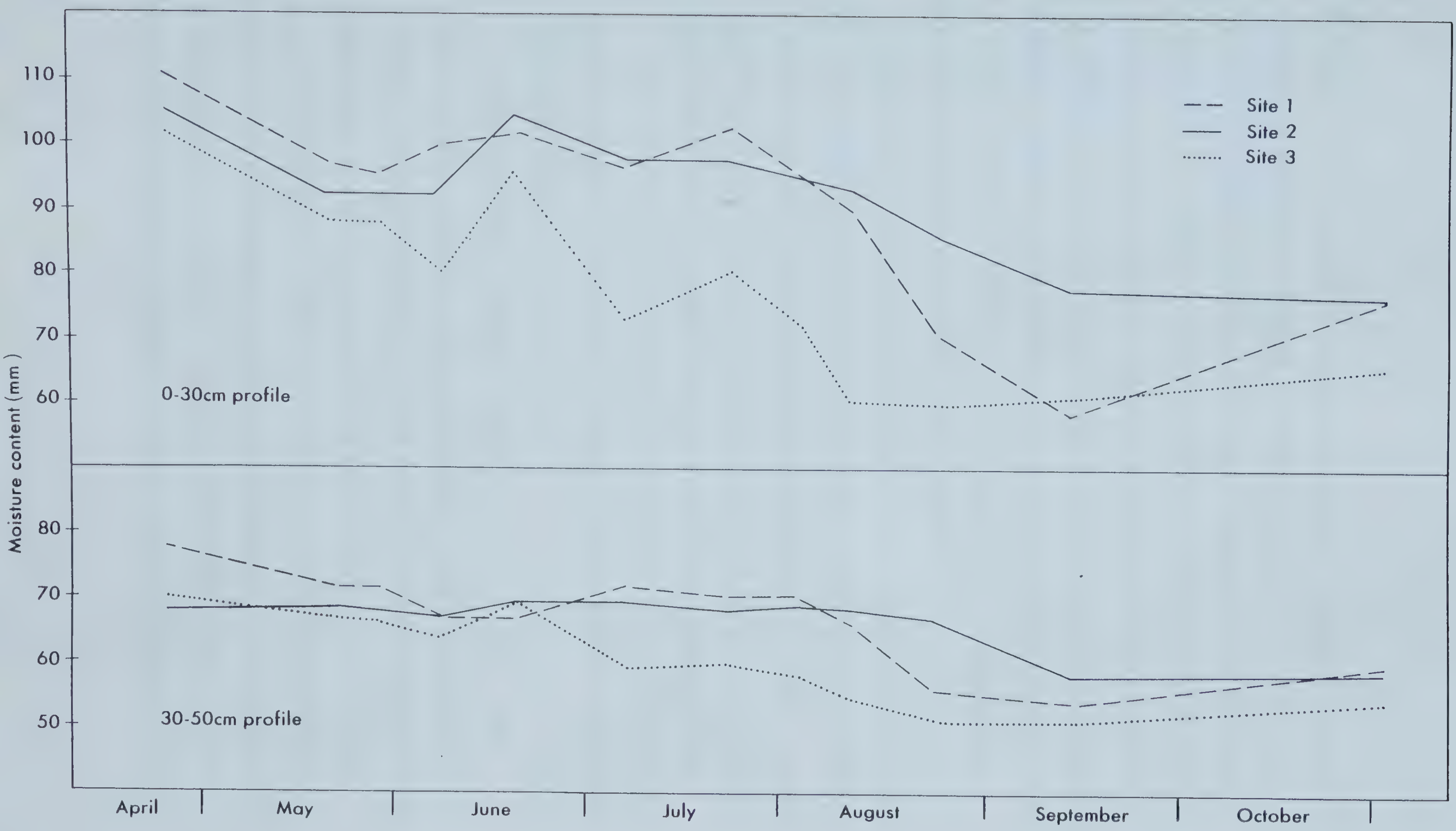

Figure 12 Soil moisture changes in the top 0-30 cm and 30-50 cm layers of the 3 Falher sites

of these soils, a result of high clay contents, may also have contributed to the lack of change by restricting water vapour loss. The fact that 1981 was a rather dry year undoubtedly contributed to the lack of changes in moisture content in the 50 to 140 cm layer as there was little moisture input into the soil at this depth. Low hydraulic conductivity also contributed to the lack of significant moisture change in these layers. Changes in moisture content recorded in the top 50 cm of the soils are a result of the proximity of this layer to plant roots and the soil surface.

The actual change in total soil moisture content varied between the three sites. Site 2 seeded with canola showed the least change of 60 mm of water between April 27 and September 14. Site 1, seeded with barley, showed the most change, decreasing 85 mm in the same period, while site 3, under alfalfa, showed a decrease of 80 mm.

The differences in profile moisture changes during the season can be attributed to a number of factors. These may be soil, crop or climatic factors.

The maximum water storage capacity of the soil may have influenced soil moisture during the season. Maximum moisture contents for the three sites to a 50 cm depth were determined by first establishing maximum _in situ_ profile moisture contents. This _in situ_ maximum moisture content was established by first ponding water on the soil surface and taking readings of wet bulk density, using the twin probe density guage, until no more change in wet bulk density occurred. Maximum moisture content in the profile

was then calculated from this maximum wet bulk density using previously determined dry profile bulk density.

The maximum moisture contents show that sites 1, 2 and 3 have total water storage capacities of 300 mm, 336 mm and 227 mm respectively for the top 50 cm of the profile. As a result site 3 would become saturated the quickest but the exact amount of water required to saturate 50 cm of soil depends on the field moisture content. Since field moisture content is usually within the available moisture range, the amount of water required for saturation would be considerable less.

Crop factors which may have influenced profile moisture include the consumptive use by the crops. Alfalfa has the highest coefficient 0.80 - 0.90 followed by barley, a small grain, 0.75 - 0.85 while canola, an oilseed, had the lowest coefficient of 0.65 - 0.75 (Hansen et al., 1980). From this it should have been expected that the alfalfa field, site 3, would lose the greatest amount of moisture during the season and site 2, the canola field the least, all other factors remaining the same.

Because of the difficulties involved in determining what proportion of each precipitation event actually contributed to soil moisture, consumptive use determinations for the different periods of the 1981 season cannot be made. Thus, only consumptive use for the month of August when no precipitation was recorded can be compared.

During this month average consumptive use for the barley crop

was 2.5 mm/day while it was 1.7 mm/day and 0.8 mm/day for the alfalfa and canola crops respectively. That barley had the highest consumptive use appears in direct contradiction of the moisture use coefficient figures given earlier, until it is noted that the barley field had a higher profile moisture content at the start of this period than did the alfalfa field. Thus the barley crop had more available moisture at the start of this period.

The cropping history of the field may partly explain differences in initial moisture contents. Field 3, being under alfalfa, is likely to have had the lowest residual moisture contents at the end of the 1980 season. This is because perennial crops use moisture right up to the period in the fall when plant growth ceases, while annual crops cease using moisture soon after maturity and harvesting. Further, perennial crops, because of their fully developed root systems, start using moisture much earlier in the season compared to annual crops that have to develop a root system each season and thus take time before their soil moisture use becomes significant.

Climatic factors that may have influenced profile moisture include evapotranspiration and precipitation. In general the potential evapotranspiration at the three sites should be similar considering they are relatively close together, with differences in actual evapotranspiration being due to different crop coefficients. Precipitation, including both snowmelt and rainfall, may, however, have been different becasuse of the areal distribution pattern of

precipitation. Precipitation may thus be a contributor to the differences in moisture between profiles.

The influence of climate on soil moisture may be interpreted from the data in Table X which includes soil moisture content, precipitation and evapotranspiration rates for the period April 27 to November 2. Soil moisture content is given for the 0 - 50 cm layer of the soil. The precipitation and evapotranspiration figures given are accumulated for each period starting the day after the soil moisture reading and ending on the day of the next moisture reading.

In general evapotranspiration increased gradually in the season from an average of 3.5 mm/day in May to a peak 6.5 mm/day in August, then decreased to 0.7 mm/day in October. The increase in evapotranspiration in the first half of the season was not, however, accompanied by a proportional decrease in profile moisture content because precipitation moderated the influence of evapotranspiration. In fact for brief periods of time during peak precipitation activity, in June and July, profile moisture actually increased.

The influence of evapotranspiration is most observable during late August and early September when no rainfall occurred. Soil moisture content registered the sharpest decline of the season during this period. One or two rain showers during late September and October are responsible for the slight increase in soil moisture content shown in the last reading of the season.

TABLE X MOISTURE CONTENT IN THE TOP 50 CM vs EVAPOTRANSPIRATION AND PRECIPITATION

Date		Potential		Site 1		Site 2		Site 3	
Month	Day	E_T (mm)	Daily E_T (mm/day)	Rain (mm)	MC (mm)	Rain (mm)	MC (mm)	Rain (mm)	MC (mm)
04	27				176		173		172
			3.5						
05	24	99		03	170	11	162	08	157
			4.5						
05	28	18		00	168	00	161	00	156
			4.5						
06	08	45		08	167	11	160	12	144
			3.5						
06	22	49		25	169	58	171	52	164
			3.6						
07	07	58		17	169	30	167	17	132
			4.9						
07	23	79		57	173	23	166	55	141
			4.8						
08	03	53		08	167	00	165	03	130
			6.4						
08	11	51		00	157	00	161	00	115
			5.4						
08	25	75		00	127	00	153	00	111
			3.0						
09	14	59			113	15	136	10	112
			0.7						
11	02	34		00	138	00	136	25	120

CONCLUSIONS

The water table does not contribute to the drainage problem in the Falher area because of its great depth which has also resulted in few seepage zones in the research area. The level landscape, however, combined with low soil hydraulic conductivity rates contributes to the drainage problem by enhancing surface ponding.

This current study has shown that drainage problems in the Falher area are compounded to a large extent by unfavourable soil physical properties. High clay contents have resulted in high water retention capacities. High clay contents coupled with high bulk densities have also resulted in the extremely low hydraulic conductivities characteristic of these soils. Different soil physical properties between the Ap and B horizons, together with the extremely low hydraulic conductivities, especially in the underlying B horizons, greatly reduce infiltration, moisture redistribution and percolation.

The high water holding capacity of these soils means that relatively small quantities of water are necessary to create saturated soil moisture conditions. The low rates of infiltration, redistribution and percolation mean that once saturated, the soils are bound to remain so until moisture is reduced through evapotranspiration. This explains why drainage problems arise during snowmelt when large quantities of water are added to the soil. The same is true during harvest when high precipitation, during

declining evapotranspiration due to lower environment temperatures may cause drainage and trafficability problems.

Whereas sites 1 and 2 show similar profile characteristics, site 3 appears to be unique. Soil moisture holding estimates show that site 3 holds significantly less water than do sites 1 and 2 at saturation. Thus site 3 requires the least amount of water to reach saturation. Furthermore, sites 1 and 2 show distinct differences in texture, water retention capacity, bulk density and saturated hydraulic conductivity between the Ap and the B horizons while differences in these properties between the Ap and the B horizon are much less pronounced at site 3.

The lack of pronounced differences in soil hydrologic properties between the Ap and the B horizons at site 3 means that water flow through the Ap is at the same rate as through the B horizon. As a result, therefore, perched water tables, a potential problem at sites 1 and 2, are less likely at site 3. The relatively lower hydraulic conductivity of the Ap horizon at site 3, however, means that infiltration rates into this site are lower and hence surface ponding is more likely.

Cultivation practices may be among the possible reasons for the lack of a pronounced difference between the Ap and the B horizon at site 3. Practices that bring to the surface high clay B material which mix with and give the Ap horizon properties similar to the B horizon may be one reason. Less disturbance of this site during the past four years that it has been under alfalfa may

have allowed the Ap to consolidate, reverting to structure similar to the B horizon and thus having properties similar to it. Management practices, both crops grown and cultivation practices, used during the period this field has been farmed may also have contributed to the similar properties between the Ap and the B horizon.

POTENTIAL SOLUTIONS TO THE DRAINAGE PROBLEM AT FALHER

DRAINAGE

One possible solution to the drainage problem in Falher is subsurface drainage. Subsurface drainage has a number of advantages over other drainage systems. Subsurface drains are installed deep enough in the soil that the drains do not interfere with field operations. Once properly installed, subsurface drains require little maintenance and so do not have the high maintenance costs of surface systems. Subsurface drains have another advantage over surface drains; they increase the root zone by effectively draining excess moisture from the soil above the depth of installation.

The design of an effective subsurface drainage system, however, is greatly dependent on the soil's hydraulic conductivity. When hydraulic conductivity is high, drains can be deeply laid and widely spaced. When hydraulic conductivities are low, drains have to be placed at shallow depths and closely spaced to be effective. With hydraulic conductivities as low as 10^{-8} cm/sec in Falher soils, drains would have to be very closely spaced, while the fact that hydraulic conductivities decrease with depth means relatively shallow laying of drains would be necessary for effectiveness. The very high cost associated with such a system may not be justifiable at Falher.

Potential silt and clay clogging of drains in these high clay and silt soils is likely to reduce a subsurface system's effectiveness and life while increasing its maintenance costs.

Based on the findings of this study, a possible solution worthy of investigation is the use of mole drains. Mole drains are particularly suited to high clay soils with low permeability in the upper layers and impervious subsoils (Raadsma, 1974). Mole drains have the advantage that they can be constructed just below the relevant root depth allowing faster drainage of the root zone.

Mole drains may have relatively short life spans of 3-15 years and may have to be closely spaced; however, because of their low cost and ease of construction (the farmer can incorporate the construction of moles in his field preparation every few years), they may be a viable venture.

Surface drains may prove useful in remedying the drainage problem in Falher because they are well suited to the rapid removal of large quantities of water (Houston, 1967). They are relatively simple to construct. To be successful, however, they require that there be some slope to the land. This may be a limitation in Falher where the land is level. The lack of natural outlets would necessitate the artifical construction of these to carry away the water, an expensive undertaking that would also be taking land out of production.

Open ditch drains also have the potential to encourage soil erosion in high silt and clay soils such as are found at Falher. Surface drains would thus have to be carefully and cautiously constructed and used.

Even though surface drains may be effective in the removal of ponded surface water, they are not effective in reducing excessive water in the soil. The removal of this excessive water from the soil is essential to solving the drainage problem completely. Surface drains would have to be coupled with some other drainage system to reduce moisture

within the soil to solve the drainage problem in Falher.

Surface drains have other disadvantages which may discourage their use. They take land out of production and also impede farm machinery movement.

DEEP PLOWING

Deep plowing is a potential solution to the drainage problem where a surface or subsurface layer of low hydraulic conductivity is the major impediment to redistribution and percolation. Under such conditions, the breaking up of this low permeability layer by deep plowing may alleviate the drainage problem.

Deep plowing of these soils may indeed increase the root zone and thus actually increase the amaount of water in the soil at saturation. This would effectively reduce soil surface ponded water and thus reduce chances of trafficability problems in these soils. However, deep plowing these soils is unlikely to increase the rate of drainage out of the profile because the water flow impeding layer extends beyond depths that can be practically deep plowed and also because the hydraulic conductivities of these soils decrease with depth.

Other problems may also arise as a result of deep plowing. High clay B horizon material may be brought up to and mixed with the surface horizon material reducing the hydraulic conductivity of the A horizon and resulting in lower infiltration rates. High clay B horizon material may also affect the tilth of the soil

making the soils more difficult to work while wet.

Though not investigated in this study, soil chemistry, especially soil salinity and the exchangeable cations, may have a profound influence on soil structure and hence on water movement through soils. Soil Na^{+} , Ca^{++} and Mg^{++} contents would have to be determined for the different horizons before deep plowing could be considered as a solution in Falher.

MANAGEMENT PRACTICES

The modification of management practices perhaps offers the most economical and practical solution to the drainage problem in Falher. Modification of cultivation practices, the time of plowing, the depth of cultivation, the change in equipment used may all help lower soil moisture. Cropping rotations and crops grown may also have an influence on soil moisture status. Cultivation practices and cropping patterns that reduce moisture should be adopted.

RECOMMENDATIONS FOR FUTURE STUDY

Considering that 1981 was an unusually dry year in the Peace River region, further investigation during an exceedingly wet year may prove instructional in evaluating the extent of the drainage problem more fully under field conditions.

The areal extent of soil physical property variability within and between fields in the study area should be investigated by monitoring more sites.

Studies should be undertaken to establish at what moisture contents trafficability becomes a problem and at what moisture contents crop growth begins to suffer. Using these moisture contents, it should then be possible to predict which precipitation events are likely to cause drainage problems. Such information is vital in the design of effective drainage systems.

The unique nature of site 3 soil physical properties raises the possibility that cultivation practices or the lack thereof may be responsible for observed physical properties of the soil. A thorough study of the effect of different management practices on the physical properties of the soil should be conducted. The effect of zero tillage and the possibility that this may result in the soil reconsolidating to its original massive structure should be investigated. The effect of different plowing depths including deep plowing should be studied as should the number of cultivation runs.

The growing of high water using perennial forages should be

investigated as this may help maintain lower profile moisture contents. Continuous growing crops in the field and eliminating fallow in the cropping rotation may be useful in maintaining lower soil moisture contents as well.

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